Contemporary Issues in Estuarine Physics

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Estuaries are areas of high socioeconomic importance, with 22 of the 32 largest cities in the world being located on river estuaries. Estuaries bring together fluxes of fresh and saline water, as well as fluvial and marine sediments, and contain many biological niches and high biological diversity. Increasing sophistication of field observation technology and numerical modeling have led to significant advances in our understanding of the physics of these systems over the last decade.

This book introduces a classification for estuaries before presenting the basic physics and hydrodynamics of estuarine circulation and the various factors that modify it in time and space. It then covers special topics at the forefront of research, such as turbulence, fronts in estuaries and continental shelves, low inflow estuaries, and implications of estuarine transport for water quality.

With contributions from some of the world’s leading authorities on estuarine and lagoon hydrodynamics, this volume provides a concise foundation for academic researchers, advanced students and coastal resource managers.

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Preface

This book resulted from the lectures of a PanAmerican Advanced Studies Institute (PASI) funded by the United States National Science Foundation and the Department of Energy. The topic of the PASI was “Contemporary Issues in Estuarine Physics, Transport and Water Quality” and was held from July 31 to August 13, 2007 at the Unidad Académica Puerto Morelos of the Mexican National University (UNAM). One of the requirements was that the PASI had to involve lecturers and students from the Americas, with most from the United States. The institute was restricted to advanced graduate students and postdoctoral participants. Because of the requirements, this book includes authors who work in the United States but tries to be comprehensive in including aspects of estuarine systems in different parts of the world. The book, however, reflects regional experiences of the authors and obviously does not include exhaustive illustrations throughout the world. Nonetheless, it is expected to motivate studies, in diverse regions, that address problems outlined herein.

This book should be appropriate for advanced undergraduate or graduate courses on estuarine and lagoon hydrodynamics. It should also serve as a reference for the professional or environmental manager in this field. The sequence of chapters is designed in such a way that the topic is introduced in terms of estuaries classification (Chapter 1). This is followed by the basic hydrodynamics that drive the typically conceived estuarine circulation consisting of fresher water moving near the surface toward the ocean and saltier water moving below in opposite direction (Chapter 2). This chapter also presents the implications of estuarine circulation on salinity stratification. The chapter sequence then deals with processes that modify the basic circulation pattern, such as tides. The theoretical framework for tides in different systems is treated in Chapter 3. The effect of tides on estuarine circulation is presented at intratidal and subtidal time scales in Chapter 4. Chapters 5 and 6 deal with effects of bathymetry on estuarine hydrodynamics. The effects of lateral bathymetry and lateral circulation on estuarine circulation are explored in
Chapter 5. Chapter 6 depicts the circulation driven by tides and winds under varying bathymetry, to compare with the results of Chapter 3 for tides. The rest of the chapters deal with selected topics related to estuarine physics: turbulence is studied in Chapter 7; fronts in estuaries and continental shelves are covered in Chapter 8; processes in low-inflow estuaries are discussed in Chapter 9; and water quality implications are presented in Chapter 10.

The effort of putting this book together was made possible by the interest and dedication of the chapter authors, whose gathering at the PASI was supported by funding from the United States National Science Foundation, under project IOISE-0614418. Special recognition to David Salas de León and Adela Monreal, from the National University of Mexico (UNAM), for their tremendous contributions and original ideas in the organization of the PASI. Mario Cáceres, David Salas Monreal, Gilberto Expósito and Miguel Angel Diaz were extremely helpful with the logistics during the PASI. Particular gratitude to the Academic Unit of UNAM in Puerto Morelos, Brigitta Van Tussembroek, Director of the Unit at the time of the PASI, for allowing the use of their facilities for this activity.
Definition and classification of estuaries

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This chapter discusses definitions and classification of estuaries. It presents both the classical and more flexible definitions of estuaries. Then it discusses separate classifications of estuaries based on water balance, geomorphology, water column stratification, and the stratification–circulation diagram – Hansen–Rattray approach and the Ekman–Kelvin numbers parameter space.

The most widely accepted definition of an estuary was proposed by Cameron and Pritchard (1963). According to their definition, an estuary is (a) a semienclosed and coastal body of water, (b) with free communication to the ocean, and (c) within which ocean water is diluted by freshwater derived from land. Freshwater entering a semienclosed basin establishes longitudinal density gradients that result in long-term surface outflow and net inflow underneath. In classical estuaries, freshwater input is the main driver of the long-term (order of months) circulation through the addition of buoyancy. The above definition of an estuary applies to temperate (classical) estuaries but is irrelevant for arid, tropical and subtropical basins. Arid basins and those forced intermittently by freshwater exhibit hydrodynamics that are consistent with those of classical estuaries and yet have little or no freshwater influence. The loss of freshwater through evaporation is the primary forcing agent in some arid systems, and causes the development of longitudinal density gradients, in analogy to temperate estuaries. Most of this book deals with temperate estuaries, but low-inflow estuaries are discussed in detail in Chapter 9.

1.1. Classification of estuaries on the basis of water balance

On the basis of the definitions above, and in terms of their water balance, estuaries can be classified as three types: positive, inverse and low-inflow estuaries (Fig. 1.1). Positive estuaries are those in which freshwater additions from river discharge, rain and ice melting exceed freshwater losses from evaporation or
freezing and establish a longitudinal density gradient. In positive estuaries, the longitudinal density gradient drives a net volume outflow to the ocean, as denoted by stronger surface outflow than near-bottom inflow, in response to the supplementary freshwater. The circulation induced by the volume of fresh water added to the basin is widely known as “estuarine” or “gravitational” circulation.

Inverse estuaries are typically found in arid regions where freshwater losses from evaporation exceed freshwater additions from precipitation. There is no or scant river discharge into these systems. They are called inverse, or negative, because the longitudinal density gradient has the opposite sign to that in positive estuaries, i.e., water density increases landward. Inverse estuaries exhibit net volume inflows associated with stronger surface inflows than near-bottom outflows. Water losses related to inverse estuaries make their flushing more sluggish than positive estuaries. Because of their relatively sluggish flushing, negative estuaries are likely more prone to water quality problems than positive estuaries.

Low-inflow estuaries also occur in regions of high evaporation rates but with a small (on the order of a few m³/s) influence from river discharge. During the dry and hot season, evaporation processes may cause a salinity maximum zone (sometimes

Figure 1.1. Types of estuaries on the basis of water balance. Low-inflow estuaries exhibit a “salt plug”.

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referred to as a salt plug, e.g., Wolanski, 1986) within these low-inflow estuaries. Seaward of this salinity maximum, the water density decreases, as in an inverse estuary. Landward of this salinity maximum, the water density decreases, as in a positive estuary. Therefore, the zone of maximum salinity acts as a barrier that precludes the seaward flushing of riverine waters and the landward intrusion of ocean waters. Because of their weak flushing in the region landward of the salinity maximum, low-inflow estuaries are also prone to water quality problems.

1.2. Classification of estuaries on the basis of geomorphology

Estuaries may be classified according to their geomorphology as coastal plain, fjord, bar-built and tectonic (Fig. 1.2; Pritchard, 1952). Coastal plain estuaries, also called drowned river valleys, are those that were formed as a result of the Pleistocene increase in sea level, starting ~15,000 years ago. Originally rivers, these estuaries formed during flooding over several millennia by rising sea levels. Their shape resembles that of present-day rivers, although much wider. They are typically wide (on the order of several kilometers) and shallow (on the order of 10 m), with large width/depth aspect ratios. Examples of these systems are Chesapeake Bay and Delaware Bay on the eastern coast of the United States.

Fjords are associated with high latitudes where glacial activity is intense. They are characterized by an elongated, deep channel with a sill. The sill is related to
a moraine of either a currently active glacier or an extinct glacier. In the sense of the glacier activity, it could be said that there are riverine and glacial fjords. Riverine fjords are related to extinct glaciers and their main source of buoyancy comes from river inputs. They are usually found equatorward of glacial fjords. Glacial fjords are found in high latitudes, poleward of riverine fjords. They are related to active glaciers and their main source of buoyancy is derived from melting of the glacier and of snow and ice in mountains nearby. Fjords are deep (several hundreds of meters) and narrow (several hundreds of meters) and have low width/depth aspect ratios with steep side walls. Fjords are found in Greenland, Alaska, British Columbia, Scandinavia, New Zealand, Antarctica and Chile.

Bar-built estuaries, originally embayments, became semienclosed because of littoral drift causing the formation of a sand bar or spit between the coast and the ocean. Some of these bars are joined to one of the headlands of a former embayment and display one small inlet (on the order of a few hundred meters) where the estuary communicates with the ocean. Some other sand bars may be detached from the coast and represent islands that result in two or more inlets that allow communication between the estuary and the ocean. In some additional cases, sand bars were formed by rising sea level. Examples of bar-built estuaries abound in subtropical regions of the Americas (e.g., North Carolina, Florida, northern Mexico) and southern Portugal.

Tectonic estuaries were formed by earthquakes or by fractures of the Earth’s crust, and creases that generated faults in regions adjacent to the ocean. Faults cause part of the crust to sink, forming a hollow basin. An estuary is formed when the basin is filled by the ocean. Examples of this type of estuary are San Francisco Bay in the United States, Manukau Harbour in New Zealand, Guaymas Bay in Mexico and some Rias in NW Spain.

1.3. Classification of estuaries on the basis of vertical structure of salinity

According to water column stratification or salinity vertical structure, estuaries can be classified as salt wedge, strongly stratified, weakly stratified or vertically mixed (Pritchard, 1955; Cameron and Pritchard, 1963). This classification considers the competition between buoyancy forcing from river discharge and mixing from tidal forcing (Fig. 1.3). Mixing from tidal forcing is proportional to the volume of oceanic water entering the estuary during every tidal cycle, which is also known as the tidal prism. Large river discharge and weak tidal forcing results in salt wedge estuaries such as the Mississippi (USA), Rio de la Plata (Argentina), Vellar (India), Ebro (Spain), Pánico (Mexico), and Itajaí-Açu (Brazil). These systems are strongly stratified during flood tides, when the ocean water intrudes in a wedge shape. Some of these systems lose their salt wedge nature during dry periods. Typical
tidally averaged salinity profiles exhibit a sharp pycnocline (or halocline), with mean flows dominated by outflow throughout most of the water column and weak inflow in a near-bottom layer. The mean flow pattern results from relatively weak mixing between the inflowing ocean water and the river water.

Moderate to large river discharge and weak to moderate tidal forcing result in strongly stratified estuaries (Fig. 1.3). These estuaries have similar stratification to salt wedge estuaries, but the stratification remains strong throughout the tidal cycle as in fjords and other deep (typically >20 m deep) estuaries. The tidally averaged salinity profiles have a well-developed pycnocline with weak vertical variations above and below the pycnocline. The mean flow exhibits well-established outflows and inflows, but the inflows are weak because of weak mixing with freshwater and weak horizontal density gradients.

Weakly stratified or partially mixed estuaries result from moderate to strong tidal forcing and weak to moderate river discharge. Many temperate estuaries, such as Chesapeake Bay, Delware Bay and James River (all in the eastern United States) fit into this category. The mean salinity profile either has a weak pycnocline or continuous stratification from surface to bottom, except near the bottom mixed layer. The mean exchange flow is most vigorous (when compared to other types of estuaries) because of the mixing between riverine and oceanic waters.

Strong tidal forcing and weak river discharge result in vertically mixed estuaries. Mean salinity profiles in mixed estuaries are practically uniform and mean flows are unidirectional with depth. In wide (and shallow) estuaries, inflows may develop on one side across the estuary and outflow on the other side, especially during the dry season. Parts of the lower Chesapeake Bay may exhibit this behavior in early autumn. In narrow well-mixed estuaries, inflow of salinity may only occur during
the flood tide because the mean flow will be seaward. Examples of this type of estuary are scarce because, under well-mixed conditions, the mean (as in the tidally averaged sense) flow will most likely be driven by wind or tidal forcing (e.g., Chapter 6).

It is essential to keep in mind that many systems may change from one type to another in consecutive tidal cycles, or from month to month, or from season to season, or from one location to another inside the same estuary. For instance, the Hudson River, in the eastern United States, changes from highly stratified during neap tides to weakly stratified during spring tides. The Columbia River, in the western United States, may be strongly stratified under weak discharge conditions and similar to a salt-wedge estuary during high discharge conditions.

### 1.4. Classification of estuaries on the basis of hydrodynamics

A widely accepted classification of estuaries was proposed by Hansen and Rattray (1966) on the basis of estuarine hydrodynamics. It is best to review this classification after acquiring a basic understanding of estuarine hydrodynamics, e.g., after Chapter 6 of this book. This classification is anchored in two hydrodynamic non-dimensional parameters: (a) the circulation parameter and (b) the stratification parameter. These parameters refer to tidally averaged and cross-sectionally averaged variables. The circulation parameter is the ratio of near-surface flow speed $u_s$ to sectionally averaged flow $U_f$. The near-surface flow speed is typically related to the river discharge and, for the sake of argument, on the order of 0.1 m/s. The depth-averaged flow $U_f$ is typically very small, tending to zero, in estuaries of vigorous water exchange because there will be as much net outflow as net inflow. In estuaries with weak net inflow, such as well-mixed and salt-wedge systems, the depth-averaged flow will be similar in magnitude to the surface outflow. Therefore, the circulation parameter is $>10$ in estuaries with vigorous gravitational circulation and close to 1 in estuaries with unidirectional net outflow. In general, the greater the circulation parameter, the stronger the gravitational circulation.

The other non-dimensional parameter, the stratification parameter, is the ratio of the top-to-bottom salinity difference $\partial S$ to the mean salinity over an estuarine cross-section $S_0$. A ratio of 1 indicates that the salinity stratification (or top-to-bottom difference) is as large as the sectional mean salinity. For instance, if an estuary shows a sectional mean salinity of 20, for it to exhibit a stratification parameter of 1 it must have a very large stratification (on the order of 20). In general, estuaries will most often have stratification parameters $<1$. The weaker the water column stratification, the smaller the stratification parameter will be.

The two parameters described above can be used to characterize the nature of salt transport in estuaries. The contribution by the diffusive portion (vs the advective
portion) of the total salt flux into the estuary can be called $\nu$. The parameter $\nu$ may oscillate between 0 and 1. When $\nu$ is close to 0, up-estuary salt transport is dominated by advection, i.e., by the gravitational circulation. In this case, mixing processes are weak, as in a highly stratified estuary (fjord). When $\nu$ approaches 1, the total salt transport is dominated by diffusive processes (e.g., tidal mixing), as in unidirectional net flows. The parameter $\nu$ may be portrayed in terms of the stratification and circulation parameters (Fig. 1.4). This diagram shows that salt transport is dominated by advective processes under high gravitational circulation or strongly stratified conditions. It also shows that diffusive processes dominate the salt flux at low circulation parameter (unidirectional net flows) regardless of the stratification parameter. Between those two extremes, the salt transport has contributions from both advective and diffusive processes. The more robust the stratification and circulation parameters, the stronger the contribution from advective processes to the total salt flux will be.

These concepts can be used to place estuaries in the parameter space of the circulation and stratification parameters. The lower-left corner of the parameter space (Fig. 1.5A) describes well-mixed estuaries with unidirectional net outflows, i.e., seaward flows with no vertical structure or type 1 estuaries. These are well-mixed estuaries, type 1a, implying strong tidal forcing and weak river discharge (or large tidal prisms relative to freshwater volumes). There are also estuaries with depth-independent seaward flow but with highly stratified conditions. These type 1b estuaries have large river discharge compared to tidal forcing. In type 1 estuaries, in general, the upstream transport of salt is overwhelmingly dominated by diffusive processes ($\nu \approx 1$, Fig. 1.5B).
Type 2 (Fig. 1.5B) estuaries are those where flow reverses at depth, and include most temperate estuaries. These systems feature well-developed gravitational circulation and exhibit contributions from advective and diffusive processes to the upstream salt transport ($0.1 < \nu < 0.9$). Type 2a estuaries are well mixed or weakly stratified and type 2b estuaries are strongly stratified. Note that strongly stratified and weakly stratified estuaries of type 2 may exhibit similar features in terms of the relative contribution from diffusive processes to the upstream salt transport (Fig. 1.5B).

Type 3 estuaries are associated with fjords, where gravitational circulation is well established: strong surface outflow and very small depth-averaged flows, typical of deep basins. This flow pattern results in large values (>100–1000) of the circulation parameter (Fig. 1.5A). Type 3a estuaries are moderately stratified and type 3b are highly stratified. The peculiarity about these systems is that the upstream transport of salt is carried out exclusively by advective processes ($\nu < 0.01$, Fig. 1.5B).

Finally, type 4 estuaries exhibit seaward flows with weak vertical structure and highly stratified conditions as in a salt-wedge estuary. In type 4 estuaries, the diffusive fraction $\nu$ lines tend to converge, which indicates that in type 4 estuaries, salt transport is produced by both advective and diffusive processes. In the Hansen–Rattray diagram, it is noteworthy that some systems will occupy different positions in the parameter space as stratification and circulation parameters change from spring to neap tides, from dry to wet seasons, and from year to year.
Analogous to the classification of estuaries in terms of the two non-dimensional parameters discussed above, estuarine systems can also be classified in terms of the lateral structure of their net exchange flows. The lateral structure may be strongly influenced by bathymetric variations and may exhibit vertically sheared net exchange flows, i.e., net outflows at the surface and near-bottom inflows (e.g., Pritchard, 1956), or laterally sheared exchange flows with outflows over shallow parts of a cross-section and inflows in the channel (e.g., Wong, 1994). The lateral structure of exchange flows may ultimately depend on the competition between Earth’s rotation (Coriolis) and frictional effects (Kasai et al., 2000), as characterized by the vertical Ekman number (Ek). But the lateral structure of exchange flows may also depend on the Kelvin number (Ke), which is the ratio of the width of the estuary to the internal radius of deformation.

The Ekman number is a non-dimensional dynamical depth of the system. Low values of Ek imply that frictional effects are restricted to a thin bottom boundary layer (weak frictional, nearly geostrophic conditions), while high values of Ek indicate that friction affects the entire water column. The lateral structure of density-driven exchange flows may be described in terms of whether the flows are vertically sheared or unidirectional in the deepest part of the cross-section (Valle-Levinson, 2008). Under low Ek (< 0.001, i.e., < −3 in the abscissa of Fig. 1.6), the lateral structure of exchange flows depends on the dynamic width of the system (Fig. 1.6). In wide systems (Ke > 2, i.e., > 0.3 in the ordinate of Fig. 1.6), outflows and inflows are separated laterally according to Earth’s rotation, i.e., the exchange flow is laterally sheared. In narrow systems (Ke < 1, i.e., <0 in the ordinate of Fig. 1.6) and low Ek (still < 0.001, i.e., < −3 in the abscissa of Fig. 1.6), exchange flows are vertically sheared. In contrast, under high Ek (> 0.3, i.e., > −0.5 in the abscissa of Fig. 1.6) and for all Ke, the density-driven exchange is laterally sheared independently of the width of the system. Finally, under intermediate Ek (0.01 < Ek < 0.1, i.e., between −2 and −1 in the abscissa of Fig. 1.6), the exchange flow is preferentially vertically sheared but exhibiting lateral variations.

The main message is that under weak friction (Ek < 0.01), both depth and width are important to determine whether the density-driven exchange is vertically or horizontally sheared. This is illustrated by the fact that the contour values in the entire region of Ek < 0.01 (i.e., <−2 in the abscissa) in Fig. 1.6 vary with both Ek and Ke. In contrast, under Ek > 0.01 the depth is the main determinant as to whether the exchange is vertically or horizontally sheared. This is shown by the fact that the contour values in Fig. 1.6 vary mostly with Ek but very little with Ke. A future challenge of this approach is to determine the variability of a particular system in the Ek vs Ke parameter space. It is likely that an estuary will describe an ellipse of variability in this plane from spring to neaps and from wet to dry seasons, or from year to year.
All of the above classifications depend on diagnostic parameters that require substantial information about the estuary, i.e., on dependent variables. In addition, they do not take into account the effects of advective accelerations, related to lateral circulation, that may be of the same order of magnitude as frictional effects (e.g., Lerczak and Geyer, 2004). Some of these nuances are discussed further in Chapter 5 of this book. Future schemes will require taking those advective effects into account. In the following chapter, in addition to presenting the basic

Figure 1.6. Classification of estuarine exchange on the basis of $Ek$ and $Ke$. The subpanels appearing around the central figure denote cross-sections, looking into the estuary, of exchange flows normalized by the maximum inflow. Inflow contours are negative and shaded. The vertical axis is non-dimensional depth from 0 to 1 and the horizontal axis is non-dimensional width, also from 0 to 1. The central figure illustrates contours of the difference between maximum outflow and maximum inflow over the deepest part of the channel and for different values of $Ek$ and $Ke$. Note that the abscissa and ordinate represent the logarithm of $Ek$ and $Ke$. Dark-shaded contour regions denote net inflow throughout the channel, i.e., laterally sheared exchange flow as portrayed by the subpanels whose arrows point to the corresponding $Ek$ and $Ke$ in the dark-shaded regions. Light contour regions illustrate vertically sheared exchange in the channel as portrayed by the subpanels whose arrows point to the corresponding $Ek$ and $Ke$ in the light-shaded regions. Intermediate-shaded regions represent vertically and horizontally sheared exchange flow, similar to the second subpanel on the left, for $\log(Ke) = 0$ and $\log(Ek) \sim -3.7$. 

All of the above classifications depend on diagnostic parameters that require substantial information about the estuary, i.e., on dependent variables. In addition, they do not take into account the effects of advective accelerations, related to lateral circulation, that may be of the same order of magnitude as frictional effects (e.g., Lerczak and Geyer, 2004). Some of these nuances are discussed further in Chapter 5 of this book. Future schemes will require taking those advective effects into account. In the following chapter, in addition to presenting the basic
hydrodynamics in estuaries, another approach for classifying estuaries based on different dynamical properties is discussed. Such an approach, consistent with that of Prandle (2009), uses only the river discharge velocity and the tidal current velocity as the parameters needed to classify estuaries.

References


2

Estuarine salinity structure and circulation

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2.1 The horizontal salinity gradient

Estuaries show a great diversity of size, shape, depth, and forcing characteristics, but a general characteristic of estuaries is the presence of a horizontal salinity gradient (Fig. 2.1). Normally the salinity decreases from the ocean toward the head of the estuary due to freshwater input; in the case of inverse estuaries, the salinity increases in the landward direction due to excess evaporation (see Chapters 1 and 9). This salinity gradient is the key dynamical variable that makes estuaries different from any other marine or lacustrine environment. The horizontal salinity gradient is the key driving force for the estuarine circulation, which in turn plays a key role in maintaining salinity stratification in estuaries. The combined influence of the estuarine circulation and stratification determines the fluxes of salt and freshwater within the estuary, and their intensity varies with the strength of the freshwater inflow. Because of these dynamics, estuaries are often the most strongly stratified aquatic environments, but they also tend to have vigorous water and salt exchange, due to the estuarine circulation.

This chapter explores the coupled equations involving the estuarine circulation, stratification, salt flux and freshwater inflow. A major outcome of this analysis is to reveal the essential importance of horizontal salinity gradient in the estuarine circulation, stratification, and salt balance, but also to find that the horizontal salinity gradient ultimately depends on the strength of the freshwater outflow and the intensity of mixing by tidal currents.

2.2. The estuarine circulation

The first known reference to the estuarine circulation was contributed by Pliny the Elder in reference to the vertically varying flow in the Strait of Bosphorus, which connects the high-salinity Mediterranean to the more brackish Black Sea. He found that as fishermen lowered their nets, they would tug in the eastward direction once
they reached a certain depth, at which they were being swept by the deep, high-salinity inflow into the Black Sea. Nineteen hundred years later, Knudsen noted a similar bidirectional flow in the entrance to the Baltic Sea. His name has been immortalized for his quantification of the salt balance, which will be discussed in Section 2.4. In 1952, Pritchard was the first researcher to link the estuarine circulation to the forcing by the horizontal density gradient, using observations in the James River estuary to demonstrate the mechanism.

Pritchard pointed out that the tidal currents are typically much stronger than the estuarine circulation, but if the vertically varying horizontal currents are measured through the course of the tidal cycle and then averaged, the “residual” or estuarine circulation would be revealed, as shown in Fig. 2.2. In a “normal” or positive estuary (see Chapter 1), i.e., one with excess freshwater input, the near-bottom flow is
landward. The strength of the estuarine circulation is typically 0.05 to 0.3 m/s (as measured by the tidally averaged bottom inflow).

The driving force of the estuarine circulation is the horizontal salinity gradient $\partial s/\partial x$, which induces a vertically varying pressure gradient. The pressure gradient $\partial p/\partial x$ can be expressed as the combined influence of the surface slope $\partial \eta/\partial x$ and $\partial s/\partial x$:

$$\frac{1}{\rho} \frac{\partial p}{\partial x} = g \frac{\partial \eta}{\partial x} + \beta g \frac{\partial s}{\partial x} (h - z),$$

(2.1)

where $\rho$ is the density of water (dominated by salinity), $\beta$ is the coefficient of saline contraction, $g$ is the acceleration due to gravity, $h$ is the water depth, and $z$ is the vertical coordinate measured upward from the bottom. Note that the second term on the right-hand side of equation (2.1), the $\partial s/\partial x$ term, is zero at the surface and maximal at the bottom, oriented in the direction that accelerates the bottom water into the estuary. The tidally averaged surface slope tilts in the other direction from the salinity gradient, with a magnitude large enough, relative to the salinity gradient, that the pressure gradient reverses somewhere close to the middle of the water column. Thus the surface water is driven seaward and the bottom water landward.

If the flow were starting from rest (imagine the estuary was frozen and it suddenly melted), the pressure gradient would result in acceleration of the surface and bottom waters in opposite directions. This acceleration would continue until some other force balanced the pressure gradient. The force that is most important for balancing the pressure gradient is the internal stress (or momentum flux) acting on the

Figure 2.2. Vertical profiles of currents in the Hudson River estuary during maximum flood, maximum ebb, tidal average, and the theoretical profile. Based on observations from Geyer et al. (2001).
estuarine shear flow. Pritchard represented the turbulent stress in terms of an eddy viscosity $A_z$,

$$\tau = \rho A_z \frac{\partial u}{\partial z}, \quad (2.2)$$

where $\tau$ is the stress (in units of force/area or Pascals) and $\partial u / \partial z$ is the vertical shear of the horizontal flow. The units of $A_z$ are length$^2$/time, just like the molecular viscosity, although it is several orders of magnitude larger. A practical way of thinking about the eddy viscosity is as a product of a turbulent velocity scale and a turbulent length scale. The magnitude of the eddy viscosity is set by the intensity of the tidal flow and the stratification; values range from $10^{-4}$ to $10^{-2}$ m$^2$/s in estuaries. The eddy viscosity varies significantly in space and time due to variations in forcing, but the momentum balance can be approximated using a constant value that represents an “effective” tidal average (although the estimate of its value is best obtained a posteriori).

The equation relating the pressure-gradient forcing to the estuarine circulation is a greatly simplified, tidally averaged representation of the horizontal momentum equation:

$$g \frac{\partial \eta}{\partial x} + \beta g \frac{\partial s}{\partial x} (h - z) = \frac{1}{\rho} \frac{\partial \tau}{\partial z} = A_z \frac{\partial^2 u}{\partial z^2}, \quad (2.3)$$

where the eddy viscosity has been assumed to be constant in the vertical to make the parameter dependence more clear. The magnitude of the surface slope is constrained by the mean outflow; if the river flow is small compared to the estuarine circulation (generally a valid approximation except for salt wedge or highly stratified estuaries), then

$$g \frac{\partial \eta}{\partial x} = -\frac{3}{8} \beta g \frac{\partial s}{\partial x} h \quad (2.4)$$

and the solution for the velocity is

$$u(z) = \frac{1}{48} \frac{\beta g \frac{\partial s}{\partial x} h^3}{A_z} \left(8\zeta^3 - 15\zeta^2 + 6\zeta \right), \quad (2.5)$$

where $\zeta = z/h$ is a non-dimensional depth varying from 0 at the bottom to 1 at the surface. The shape of the velocity profile is shown in Fig. 2.2. This should be considered a qualitative solution – the vertical structure of the velocity in actual estuaries differs due to the spatial and temporal variations of the eddy viscosity as well as other factors such as lateral advection (note comparison with actual observed profiles in Fig. 2.2). Nevertheless, equation (2.5) yields the general shape and magnitude of the estuarine circulation, and it is a useful approximation for analyzing the influence of the estuarine circulation on the stratification and salt balance.
In practice it is difficult to assign an appropriate value to $A_z$ without actually solving the time-dependent equations with the inclusion of an advanced turbulence closure. A simple alternative to equation (2.5) can be obtained by replacing $A_z$ with an equivalent expression involving the tidal velocity magnitude $U_t$, the depth $h$, and a bottom drag coefficient $C_d$: 

$$A_z = \frac{1}{48 \ a_o} C_d U_t h,$$  

(2.6)

where the drag coefficient equates the magnitude of the bottom stress to the tidal velocity as $\tau_b = \rho C_d U_t^2$ (with typical values of $C_d \sim 3 \times 10^{-3}$) and $a_o$ is a dimensionless constant yet to be determined related to the effectiveness of turbulent momentum flux. Note that this is an “effective” viscosity based on the tidally averaged flow, and its value is significantly smaller than the maximum eddy viscosity that would occur within a tidal cycle. Substituting this into equation (2.5), the magnitude of the estuarine circulation can be determined as 

$$U_e = a_o \beta g \frac{\partial s}{\partial x} \frac{h^2}{C_d U_t}.$$  

(2.7)

This formulation indicates the parameter dependence of the estuarine circulation relative to the key variables: linearly with $\partial s/\partial x$, inversely with tidal velocity, and quadratically with depth. Comparison with data from the Hudson River estuary verifies that equation (2.7) provides a reasonable estimate of the variability, with a value of $a_o$ of about 0.3 (Fig. 2.2).

### 2.3. The stratification

The vertical variation of salinity, or stratification, is one of the more conspicuous characteristics of estuaries. Estuarine stratification varies considerably from one estuary to another, and one way of classifying estuaries (Chapter 1) is based on the strength of stratification: well mixed, partially mixed, highly stratified, and salt wedge. As a classification scheme this is not very reliable, however, as one estuary may vary from well mixed to highly stratified depending on the forcing conditions. Estuarine stratification is important for a number of reasons – it inhibits vertical mixing, which affects the dynamics and may lead to hypoxia in the subpynocline waters. The stratification also plays a fundamental role in the salt balance, which will be discussed in Section 2.4.

What controls the stratification in estuaries? What makes stratification vary so much from one estuary to another, and for one estuary to vary from well mixed to highly stratified? The key to the variability of stratification is embodied in the
equation for local salt conservation, which can be simplified for steady-state conditions and modest along-estuary variations in depth to

\[ u(z) \frac{\partial s}{\partial x} = K_z \frac{\partial^2 s}{\partial z^2}, \]  

(2.8)

where \( u(z) \) is the vertically varying, tidally averaged velocity, and \( K_z \) is the eddy diffusivity of salt, with the same dimensions and similar magnitude to \( A_z \). It is not immediately obvious why this equation would dictate the stratification. One way to illustrate the balance more clearly is to integrate equation (2.8) from the middle of the water column to the surface (neglecting vertical variations in \( \partial s/\partial x \)), obtaining

\[ \alpha U_e \frac{\partial s}{\partial x} = -K_z \frac{\partial s}{\partial z_{mid}}, \]  

(2.9)

where \( \alpha \) is a constant of integration approximately equal to 0.3, and \( \partial s/\partial z_{mid} \) is the vertical gradient of salinity in the middle of the water column. This equation indicates that mean advection of salt by the estuarine circulation is balanced by vertical mixing. The salinity of the upper layer would decrease due to seaward advection of lower-salinity water, but vertical mixing carries higher-salinity water up, balancing the effect of advection (Fig. 2.3). In the lower layer, advection would cause the salinity to increase, but vertical mixing transports low-salinity water downward to maintain a steady state.

As with the eddy viscosity, the eddy diffusivity can be represented in terms of tidal velocity and depth [cf. equation (2.6)]. Substituting this into equation (2.9) and noting that \( \partial s/\partial z \approx \Delta s/h \), where \( \Delta s \) is the top-to-bottom salinity difference, we obtain

\[ \Delta s = a_1 \frac{U_e h}{C_D U_t}, \]  

(2.10)

Figure 2.3. Schematic cross-section of an estuary showing the influence of advection and vertical mixing on the local salt balance. The thick lines are isohalines, and the thin lines and arrows indicate the estuarine circulation. In Box A (upper layer), horizontal advection causes a reduction of salinity, but vertical mixing compensates by replacing low-salinity water with underlying high-salinity water. The relative roles of advection and mixing are reversed in Box B (lower layer).
where $a_1$ is a constant that incorporates $\alpha$ and also depends on the shape of the salinity and velocity profiles. The value of $a_1$ was estimated from Hudson River data to be approximately 50. Combining equation (2.10) with equation (2.7) for the estuarine circulation,

$$\Delta s = a_0 a_1 \frac{\beta g \left( \frac{\partial s}{\partial x} \right)^2 h^3}{(C_D U_t)^2}.$$  \hspace{1cm} (2.11)

This equation indicates that the stratification depends quadratically on the salinity gradient and inversely on the square of the tidal velocity — i.e., the stratification is a lot more sensitive to changes in these forcing variables than is the estuarine circulation.

The sensitive dependence of stratification on the strength of the tides has been revealed in a number of studies of the spring–neap variations in stratification. During spring tides, tidal mixing is maximal, and equation (2.11) predicts that the stratification should reach a minimum, and vice versa. Haas (1977) first documented the spring–neap variations of stratification due to variations in tidal mixing in the subestuaries of Chesapeake Bay.

Data from the Hudson estuary are used to test equation (2.11), as shown in Fig. 2.4 (lower panel). These data indicate that the actual stratification is more sensitive to the spring–neap cycle than equation (2.11) predicts. The reason for this is that the

![Figure 2.4](image_url)

Figure 2.4. Time series observations from the Hudson River in the fall of 1995 showing fortnightly and monthly variations in tidal velocity amplitude (upper panel), estuarine velocity (middle panel), and stratification (lower panel). The theoretical predictions for estuarine velocity and stratification are also shown, based on observed values of $U_T$ and $\partial s/\partial x$. 

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intensity of mixing does not just depend on the intensity of the tides; it also depends on the ambient stratification. This is a complex topic, which is treated in more detail in Chapter 7. Briefly, the parameter that best quantifies the importance of stratification on mixing is the gradient Richardson number $Ri$ – the ratio of the stratification to the shear. This can be approximated for scaling purposes by a non-dimensional ratio of the vertical salinity difference to the tidal velocity:

$$Ri_T = \frac{\beta g \Delta sh}{U_T^2}. \quad (2.12)$$

For $Ri_T > 1$, mixing is strongly suppressed by the stratification, and for $Ri_T < 0.25$, mixing is relatively unaffected by stratification. Models of mixing in stratified boundary layers (e.g., Trowbridge, 1992) suggest that the mixing rate should depend on $Ri^{-1/2}$. If equation (2.11) is modified by the addition of a factor to account for the $Ri$-dependence,

$$\Delta s = a_o a_1' Ri_T^{1/2} \frac{\beta g \left( \frac{\partial s}{\partial x} \right)^2 h^3}{(C_D U_T)^2}, \quad (2.13)$$

the fit with the Hudson River data is significantly improved (Fig. 2.4). An explicit form for $\Delta s$ that includes the stratification effect on turbulence is obtained by combining equations (2.12) and (2.13):

$$\Delta s = a_o a_1'^2 \frac{Ri_T}{(C_D U_T)^2} \left( \frac{\partial s}{\partial x} \right)^4 h^7, \quad (2.14)$$

Note the extreme sensitivity of the stratification to the tidal mixing according to this formulation, as well as the sensitivity to the horizontal salinity gradient and depth. The Hudson data provide some support for the sensitivity to tidal velocity, but equation (2.14) appears to overemphasize the dependence on horizontal salinity gradient, based on observations and model results. The consequences of this stratification-dependence on the overall estuarine balance are discussed in Section 2.5.

### 2.4. The salt balance

If an estuary is in steady state, then the amount of salt being transported past any cross-section has to be zero (except in the highly unusual case that there are significant sources of salt in the watershed). Considering a cross-section within the estuary with some tidally averaged value of salinity $s_o$, the freshwater outflow $Q_r$ (volume per time) will cause a seaward transport $Q_{salt} = Q_r s_o$. In order to maintain a steady state, there must be mechanisms that transport salt into the estuary, to
compensate for the advective loss due to the river outflow. Knudsen’s analysis of the estuarine circulation of the Baltic Sea provides the essence of the estuarine salt balance. Knudsen’s relation considers the integrated conservation of volume and salt in a basin that has a riverine source of freshwater and exchange flow with the ocean (Fig. 2.5). Volume and salt conservation yield

\[ Q_1 = Q_2 + Q_r, \]
\[ Q_1s_1 = Q_2s_2, \]  

(2.15)

where \( Q_1 \) and \( Q_2 \) are the volume transports in the upper and lower layers (the upper layer being directed seaward and the lower layer landward). Combining the two leads to Knudsen’s relation:

\[ Q_r s_1 = Q_2 \Delta s. \]  

(2.16)

If \( Q_2 \gg Q_r \) (which is generally the case for partially mixed and well-mixed estuaries), then \( s_1 \approx s_2 \approx s_o \), and Knudsen’s relation can be restated using \( U_e \):

\[ U_e s_o = a_2 U_e \Delta s, \]  

(2.17)

where \( u_r = Q_r / A_{cs} \) (the outflow velocity associated with the river discharge, \( A_{cs} \) being the local cross-sectional area of the estuary), and \( a_2 \) is a constant equal to approximately 0.5. Equation (2.17) indicates that the tendency for salt to be carried out of the estuary by the freshwater outflow is balanced by the net input of salt due to the estuarine circulation. The stronger the estuarine circulation, and the larger the vertical salinity difference, the more salt is transported into the estuary.

Equation (2.17) represents the steady-state balance, and it only considers the influence of the estuarine circulation on the salt balance. The salt content of estuaries is not actually constant in time, because a number of factors cause the left-hand and right-hand sides of equation (2.16) to change. The magnitude and causes of those variations will be discussed in Section 2.6. This equation also leaves out another contributor to the horizontal salt flux: the horizontal dispersion of salt (tidal dispersion), which is due mainly to tidal stirring. Tidal dispersion is particularly important
in short estuaries, in which the length of the salt intrusion is comparable to the tidal excursion distance (5–15 km), and also in regions of abrupt changes in estuarine cross-section. But in large estuaries (>30 km long), and away from major changes in cross-sectional geometry, the salt transport induced by estuarine circulation generally dominates over tidal dispersion, and equation (2.17) is a good representation of the time-averaged salt balance.

2.5. The coupled equations

The key to the dynamics of estuaries is that the global salt balance has to be satisfied at the same time that the local momentum and salinity equations are in balance. This combination of equations for estuarine circulation, stratification, and salt balance leads to a constraint on the horizontal salinity gradient, which is in essence the master variable controlling the estuarine dynamics. First we will consider the formulation in which the influence of stratification on mixing is neglected, as it leads to a result that is often noted in the literature. Combining equations (2.7), (2.11) and (2.17), we obtain

\[ U_r s_o = a_o^2 a_1 a_2 \frac{(\beta g)^2 (\frac{\partial s}{\partial x})^3}{(C_d U_t)^3} h^5. \]  

(2.18)

Then, solving for \( \frac{\partial s}{\partial x} \), we obtain

\[ \frac{\partial s}{\partial x} = \frac{1}{(a_o^2 a_1 a_2)^{1/3}} \frac{C_d U_t^{1/3} s_o^{1/3}}{(\beta g)^{2/3} h^{5/3}}. \]  

(2.19)

Equation (2.19) indicates that for steady-state conditions, the horizontal salinity gradient depends on the one-third power of the river flow and the first power of the tidal velocity. These scaling relations are consistent with the Hansen and Rattray (1965) solution for the advection-dominated limit, the Chatwin (1976) solution for partially mixed estuaries, and the MacCready (1999) solution for advection-dominated estuaries. The essential finding is that the salinity gradient is relatively insensitive to variations in river flow, or alternatively that the estuarine salt flux is particularly sensitive to the salinity gradient.

Now, considering the solution for the horizontal salinity gradient when the Richardson number dependence on vertical mixing is included, we start with equation (2.14) for \( \Delta s \) in combination with equations (2.7) and (2.17) and obtain

\[ \frac{\partial s}{\partial x} = \frac{1}{(a_o^3 a_1^2 a_2)^{1/5}} \frac{C_d U_t^{7/5} U_r^{1/5} s_o^{1/5}}{(\beta g)^{4/5} h^{9/5}}. \]  

(2.20)
When the influence of vertical mixing is included, the salinity gradient is even less sensitive to variability of river flow, i.e., the estuary is “stiff” (think of a steel spring) with respect to variations in river flow. Monismith et al. (2002) found even greater “stiffness” in the salinity gradient of northern San Francisco Bay, obtaining a $U_r^{1/7}$ dependence. This may be due in part to variations in geometry along the estuary, but the influence of stratification on mixing is likely an important contributor to the estuarine response to variations in river flow.

The expression for \( \partial s/\partial x \) can be substituted back into the equations for \( U_e \) and \( \Delta s \) to obtain expressions for the estuarine circulation and stratification in terms of the forcing variables \( U_t \) and \( U_r \), first without considering the Richardson number effect:

\[
U_e = \left( \frac{a_o}{a_1a_2} \right)^{1/3} \beta g s_o h^{1/3} U_r^{1/3},
\]

\[
\frac{\Delta s}{s_o} = \left( \frac{a_1}{a_oa_2^2} \right)^{1/3} \beta g s_o h^{1/3}.
\]

These equations reveal the surprising result that for steady-state balances, the estuarine circulation and stratification do not depend on the tidal velocity, even though equations (2.7) and (2.11) clearly indicate the inverse dependence on tidal mixing. This paradoxical result is due to the variation of horizontal salinity gradient with tidal velocity [equation (2.19)], which exactly compensates for the variations in tidal mixing. Real estuaries do show large variations due to changes in tidal mixing – this is because the steady-state assumption is violated in the spring–neap cycle – more on this in the next section.

For the case in which Richardson number effects are considered, we get slightly different equations for \( U_e \) and \( \Delta s \):

\[
U_e = \left( \frac{a_o^2}{a_1^2a_2} \right)^{1/5} (\beta gs_o h)^{1/5} U_r^{1/5} U_T^{2/5},
\]

\[
\frac{\Delta s}{s_o} = \left( \frac{a_1}{a_oa_2} \right)^{2/5} \frac{U_r^{4/5}}{U_T^{2/5} (\beta gs_o h)^{1/5}}.
\]

These equations indicate that the estuarine velocity \( U_e \) increases with increasing river flow (as expected), and also increases with increasing tidal velocity (not expected). Stratification is found to be slightly more sensitive to river flow than in the case without the Richardson number dependence, and also it is found to vary inversely with tidal velocity.

Although the physics would suggest that equations (2.23) and (2.24) provide a more realistic representation of the variability of estuarine circulation and
stratification, there have not been adequate analyses of these quantities among different estuaries to provide a definitive assessment. An analysis of stratification among a wide variety of estuaries (Fig. 2.6) indicates that the simple power law prediction of equation (2.22) does a better job than equation (2.24) of predicting the wide range of stratification observed in estuaries, even though it has a more simplified theoretical basis. There are other factors, such as estuarine geometry and temporal variability, that are not included in the theory but contribute to the actual variability among estuaries. These equations should be regarded at this time as providing guidance for the interpretation of variability among estuaries and within a particular estuary.

2.6. Temporal variability of the estuarine salinity structure and circulation

The combined equations presented in the previous section are based on a steady-state salt balance, meaning that the estuary neither gains nor loses salt. An estimate of the time scale for which that assumption is valid can be approximated by the
flushing time scale, which can be estimated as the ratio of the length of the salinity intrusion $L$ to the estuarine velocity:

$$T_F = \frac{L}{U_E}. \quad (2.25)$$

For the Hudson estuary, that time scale is approximately 10 days; for the Chesapeake, closer to 30; and for the Columbia River estuary, close to 1 day. If the forcing occurs at time scales comparable to or less than the residence time, then the salt balance will not “keep up” with the change in forcing, and so $\partial s/\partial x$ will adjust to the average forcing conditions rather than their short-term variations. This does not invalidate the coupled equations, but it means that they should be considered a representation of the conditions averaged over the flushing time scale.

A major contributor to temporal variability in estuaries is the fortnightly variation of tidal amplitude, mentioned earlier in the context of stratification variations. For larger estuaries such as the Hudson or the Chesapeake, $\partial s/\partial x$ remains nearly constant through the fortnightly cycle, so the variation of stratification is not predicted by the coupled equations [equation (2.22) or equation (2.24)], but rather by the “local” stratification balance [equation (2.11) or equation (2.14)]. Indeed, the large spring–neap variability of the stratification in the Hudson estuary (Fig. 2.4) indicates a significant amplification of the tidal variation, consistent with equation (2.24). In estuaries that are shorter and have faster response times, $\partial s/\partial x$ will increase during spring tides, partially offsetting the influence of increased mixing.

Other factors, such as changes in river flow and wind forcing, also contribute to temporal variability in estuaries, with the same caveat that they may occur at time scales faster or slower than the response time of the estuary, and their influence on the estuarine regime will depend on the relative time scales. The influence of temporal variability (including tides as well as lower-frequency processes) on the time-averaged estuarine regime is one of the most important topics in estuarine physics. Chapters 4, 5 and 9 examine these time-dependent processes.

### 2.7. Estuarine classification

Hansen and Rattray (1966) developed a scheme for estuarine classification called a “stratification–circulation” diagram, which has been the most commonly used approach over the last four decades. The Hansen–Rattray approach is intended as a diagnostic tool – given the observed stratification and circulation, what are the processes responsible for salt flux? Another approach to classification is the prognostic approach, in which the estuary is classified based on forcing variables, and the purpose of the classification is to predict the estuarine regime based on those forcing conditions. Due to the complexity of the processes and variability within and among estuaries, the prognostic approach could at best provide a rough estimate of the
conditions of a particular estuary – the quantitative prediction of estuarine processes is difficult even with a high-resolution numerical model. Nevertheless, the coupled equations provide a starting point for a prognostic classification of estuaries. This analysis suggests that the freshwater velocity $U_R$ and tidal velocity $U_T$ are “master variables” that may provide the framework for such a prognostic approach, as illustrated in Fig. 2.7. More strongly stratified estuaries appear on the upper part of the diagram, and weakly stratified on the lower part. The length and flushing times depend on both parameters. Additional research may lead to a more quantitative approach using this framework.

Figure 2.7. A framework for prognostic estuarine classification. The axes are the forcing variables: tidal velocity and “freshwater velocity”, non-dimensionalized by a densimetric velocity scale. Stratification variations are mainly represented by the vertical position on the diagram. The length and flushing times depend on both parameters. Additional research may lead to a more quantitative approach using this framework.
what other variables should be considered to yield a reasonable prediction of the estuarine regime.

References


