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2.2.2 LAKES

Although lakes are distinguished from rivers in part by the relative absence of a pronounced downstream flow of water, the waters of a lake are by no means stationary. Water currents, typically driven by wind instead of gravity, are a major feature of these water bodies. Water currents not only provide advective transport of chemicals in lakes but also cause transport by eddy diffusion because the water currents are almost always turbulent. In a lake, the average amount of time that water remains in the lake is called the *hydraulic residence time,* which can be estimated by the ratio of the lake volume to the rate at which water is lost through all processes (e.g., outflow, seepage, and evaporation).

TABLE 2-3 Reported Longitudinal Dispersion Coefficients[a]

River	Depth (m)	Width (m)	Velocity (m/sec)	Longitudinal dispersion coefficient (m^2/sec)
Irrigation canal	0.14	1.5	0.33	1.9
Monocacy	0.32	35	0.21	4.7
Monocacy	0.45	37	0.32	13.9
Monocacy	0.88	48	0.44	37.2
Yadkin	2.33	70	0.43	111
Yadkin	3.85	72	0.76	260
Susquehanna	1.35	203	0.39	92.9
Sabine	2.04	104	0.58	316
Sabine	4.75	128	0.64	670
Missouri	2.70	200	1.55	1500

[a] Rutherford (1994).

Wind-Driven Advection

Figure 2-6 a shows the simplest pattern of water movement in a lake, caused by wind exerting a force on the water at the lake surface. The downwind surface current is called *wind drift* and typically moves at a rate of 2 to 3% of the average wind speed. Clearly, the water moving downwind cannot pile up indefinitely at the end of the lake; instead, it flows back upwind, typically at

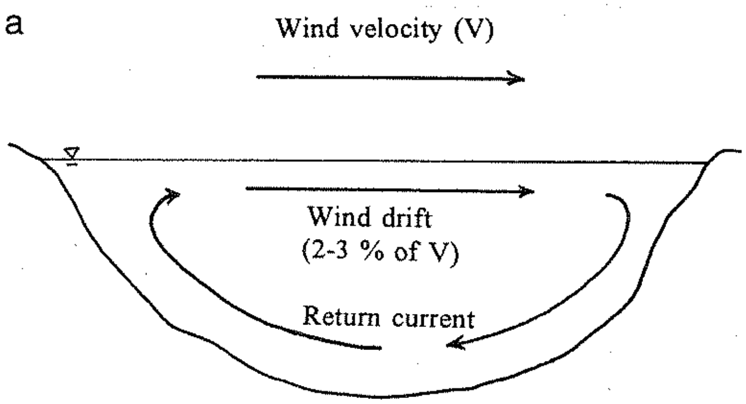

FIGURE 2-6 Wind-driven currents in a lake. (a) Circulation in a small lake of simple geometry. The surface water current, or *wind drift*, averages 2 to 3% of wind velocity. Deeper in the lake, a *return current* is established, returning water to the upwind end of the lake. (b) In a large lake such as Lake Michigan, variability of winds, complex lake geometry, and other forces (such as the Coriolis effect) lead to complex patterns of water movement [Ayers *et al.* (1958)] *(Figure continues)*.

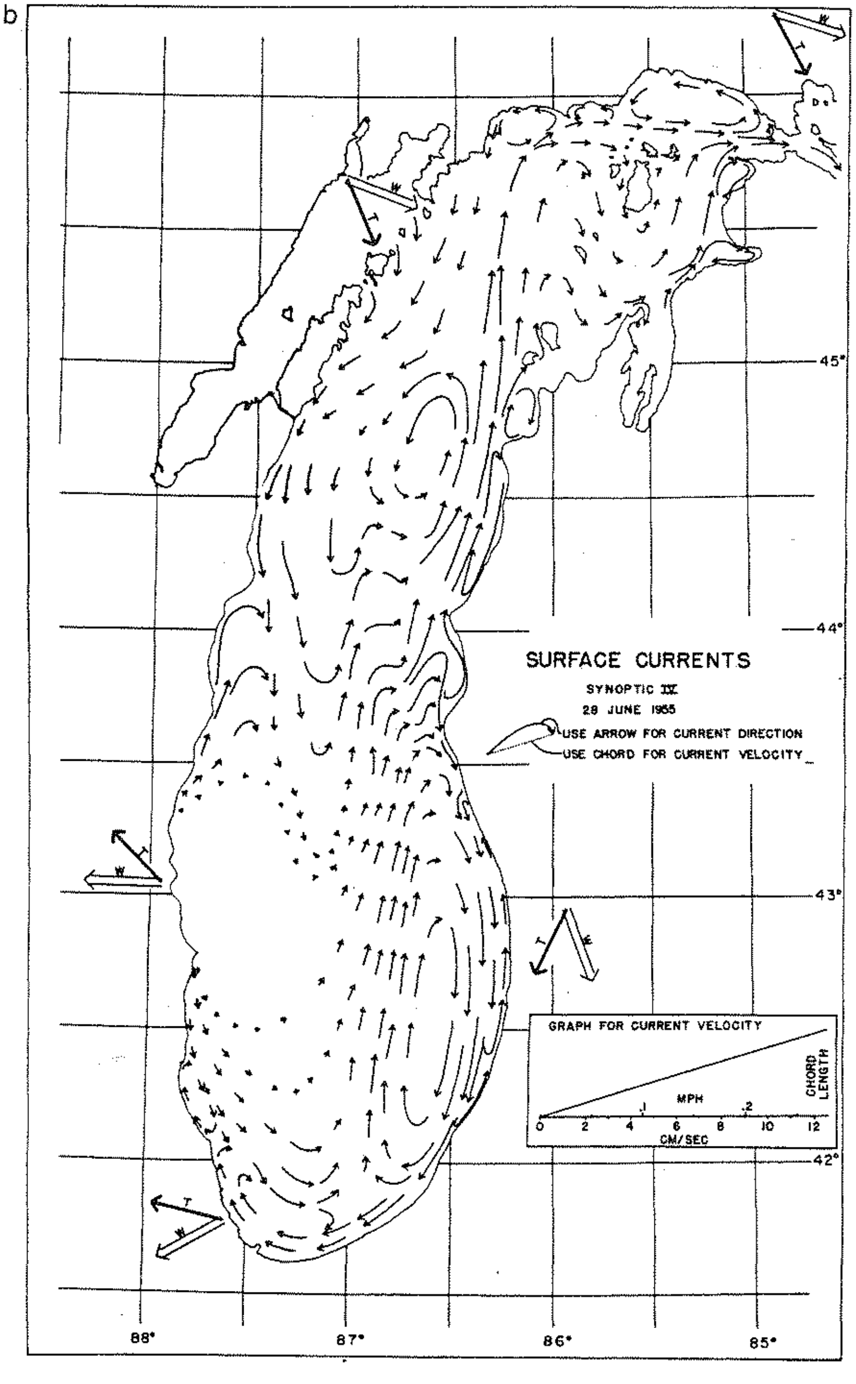

FIGURE 2-6 *(Continued)*

a greater depth, as a *return current.* Chemicals dissolved or suspended in the water are thereby transported within the lake.

In many lakes, the pattern of water movement is much more complicated, because the water movement also is affected by the shape of the lake basin, by variations in water density, by inflowing streams, and, especially in large lakes, by the Coriolis force (see Section 4.3.1). In large lakes of complex shape, the actual water currents can be exceedingly complicated, as shown in Fig. 2-6 b.

Fickian Mixing

A mass of tracer chemical injected into a lake not only will move by advection associated with water currents, but also will spread out into an ever-larger volume of water. Given enough time (perhaps a few days for a small lake), a tracer tends to become completely mixed throughout a lake; concentrations become essentially homogeneous (the same everywhere), and therefore concentration gradients become zero. Such mixing is due primarily to turbulence; eddies carry chemicals away from regions of relatively high concentration toward regions of lower concentration. The situation is comparable to that in a river, except that a one-dimensional description of Fickian mixing is rarely an adequate approximation in a lake. Lateral dispersion must almost always be taken into account because a lake's width is a significant fraction of its length, and a simplifying assumption of complete mixing across the width is not justified. If the lake is much longer and wider than it is deep (as most lakes are), and if one is mostly interested in horizontal chemical transport, it may be sufficient to consider Fickian mixing coefficients along two horizontal directions, while assuming the chemical of interest is well mixed vertically within the lake or within a stratified layer of the lake. For small time intervals after a release or in locations where the water is quite deep, however, the rate of vertical mixing must be evaluated as well. Note that Fickian mixing coefficients are generally different in each direction, just as they are for a river.

A useful two-dimensional expression for concentration of a chemical introduced as a pulse over the depth of a vertically mixed layer of water is given by Eq. [2-18]. (The equation includes the effect of first-order decay.) This equation is a solution to Eq. [1-6] under conditions of an instantaneous injection of mass into an infinite two-dimensional body of water,

$$C(x, y, t) = \frac{M}{4\pi t\sqrt{D_x D_y}}\, e^{-((x-V_x t)^2/4D_x t+(y-V_y t)^2/4D_y t)} \cdot e^{-kt}, \qquad [2\text{-}18]$$

where C is the concentration of tracer chemical [M/L^3], M is the mass of tracer chemical per depth of water [M/L], x and y are the distances from injection location along the x and y axes [L], t is the time elapsed since injection [T], V_x is the average velocity in the x direction [L/T], V_y is the average velocity in the y direction [L/T], D_x is the Fickian transport coefficient in the x direction [L^2/T], D_y is the Fickian transport coefficient in the y direction [L^2/T], and k is the first-order decay rate constant [T^{-1}]. Note that the depth of water could correspond to the total depth of a vertically well-mixed lake or to the depth of a particular layer in a stratified lake. As in Eqs. [2-10] and [2-11], simplification may be made for ideal tracer chemicals; in the absence of decay, the factor e^{-kt} becomes one.

Estimating D_x and D_y in a lake is more complicated than estimating Fickian mixing coefficients in a river, in part because of the larger areal extent of a lake, which leads to a scale issue in conducting tracer experiments. The eddies that exist in a lake may become larger than those in a river; the relatively smaller width of a river tends to limit the size of the largest eddies, while in a lake the largest eddies may be a sizable fraction of the lake basin size. The practical result is that a dispersion coefficient that is determined for a small plume of spreading chemical may not be appropriate to accurately model chemical dispersion at a later time, when the plume has spread and larger eddies are contributing to its mixing. Often the values of D are approximately proportional to the distance a plume has traveled raised to the 4/3 power. For more information on the estimation of dispersion coefficients and the use of lake models in which D_x and D_y increase with the areal extent of a chemical plume, the reader is referred to Fischer *et al.*, 1979.

$D_2 \cong D_1 \cdot L^{4/3}$

Hydrostatic Pressure

In a surface water body, water pressure can be very closely approximated by the expression,

$$P = \rho g z, \qquad [2\text{-}19]$$

where P is the water pressure [M/LT^2], ρ is the density of water [M/L^3], g is acceleration due to gravity [L/T^2], and z is the depth below the water surface [L]. The proportionality between pressure and depth is a direct result of the fact that water is nearly incompressible, and hence has a nearly constant density. As will be seen in Section 4.1.1, the compressibility of air results in a nonlinear pressure–height relationship in the atmosphere.

Stratification

Although water density does not vary greatly from 1.0 g/cm^3, the small changes in density that do occur due to variations in water temperature and solute content can have profound effects on mixing processes in a surface

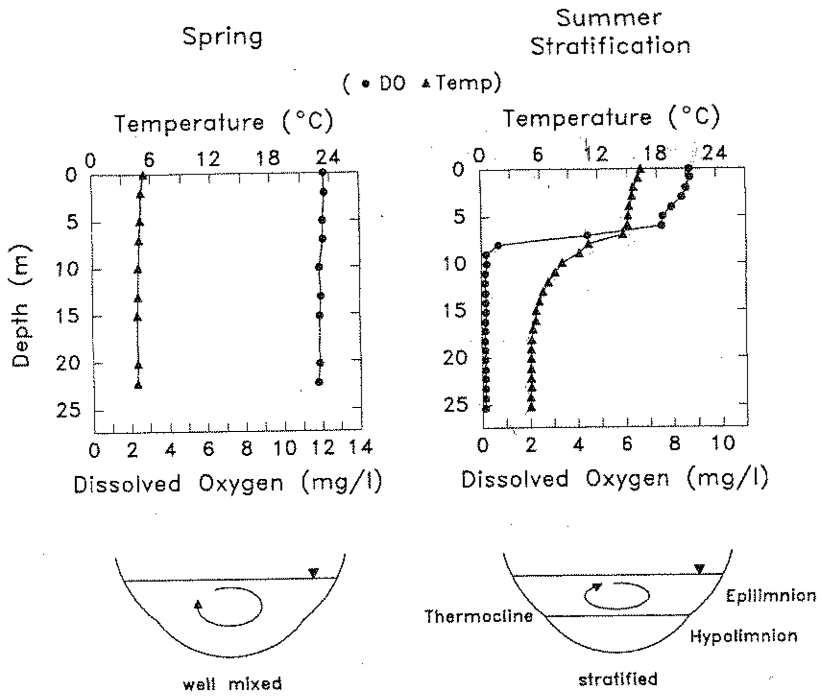

FIGURE 2-7 Measured temperature and oxygen profiles from the Upper Mystic Lake in eastern Massachusetts, on April 1, 1991 and September 30, 1991. (Left) the lake is unstratified and well mixed during turnover, which occurs in spring and fall. (Right) during summer, this eutrophic (productive) lake becomes depleted in oxygen in the lower layer of water (the hypolimnion), while its upper layer (epilimnion) remains well mixed by the wind and oxygenated by photosynthesis and by contact with the atmosphere. An oligotrophic (unproductive) lake may retain its high springtime concentration of oxygen in the hypolimnion throughout the summer [data from Aurilio (1992)].

water body. As alluded to earlier, *stratification* divides lakes into different layers by inhibiting vertical mixing between the layers. Stratification occurs when water at the bottom of a lake is denser than the surface water, and water currents (usually wind-driven) fail to generate eddies strong enough to penetrate the boundary between the water layers. Wind-driven circulation and turbulent mixing in such a lake are thereby restricted to the upper water layer (Fig. 2-7); the lower layer, isolated from wind effects, may be quite quiescent. Such a density difference is usually due to temperature differences between upper and lower water masses; the lake is then called *thermally* stratified. The upper layer, which is typically well mixed, is called the *epilimnion;* the lower layer is the *hypolimnion;* and the region separating them is the *thermocline.*

Thermal stratification is common in lakes located in climates with distinct warm and cold seasons. Although many variations are possible, the classic temperate zone lake begins a period of summer stratification when heat from solar radiation preferentially warms the uppermost water, decreasing its den-

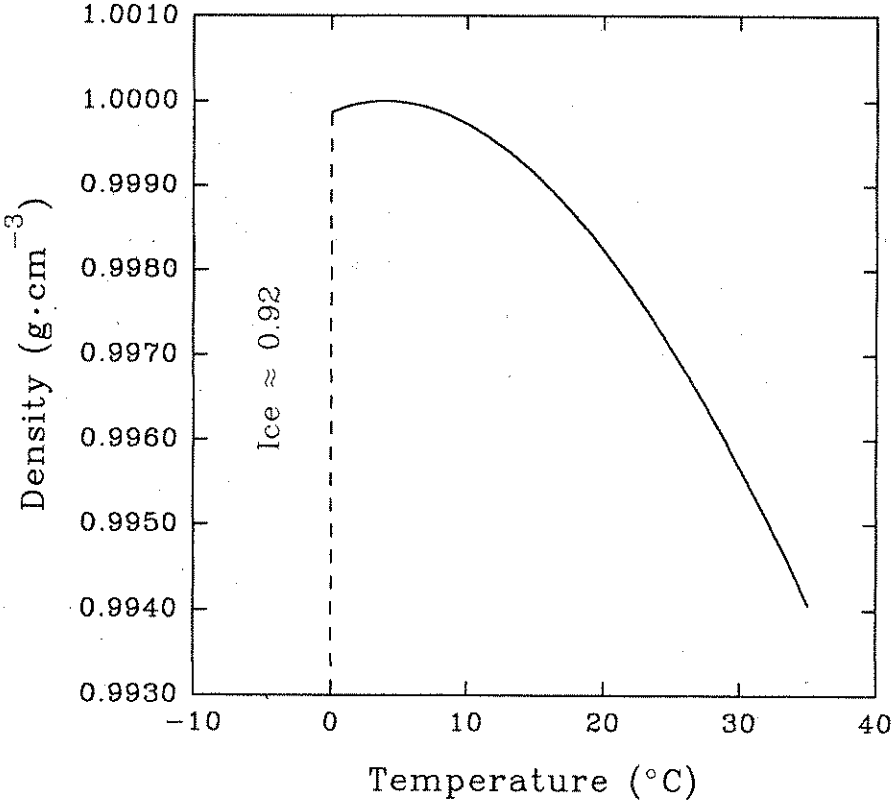

FIGURE 2-8 Density versus temperature curve for water. Maximum density occurs at 4°C; thus, stratification in a lake can occur in winter with bottom waters near 4°C and less dense surface waters closer to 0°C. In summer, if stratification occurs, the warmer water will be at the surface. Note that a given spread in water temperature conveys a larger density contrast between the waters (and hence a more stable stratification) at higher temperatures than at lower temperatures. The density of ice is *much* less than the density of liquid water (note the broken scale for ice density).

sity relative to deeper water. Usually this occurs in the spring during a few days of bright sunlight and low winds. Once started, the thermal stratification usually persists until the autumn; typical wind forces usually do not remix a lake once two layers of significantly differing densities have been created. The epilimnion does, however, thicken and become warmer throughout the summer as more solar radiation is absorbed. Remarkably little exchange of heat or chemicals occurs between the epilimnion and the hypolimnion; the lake does not fully mix again until diminished solar radiation in the autumn causes the epilimnion to cool to the approximate temperature of the hypolimnion. At this time, when the lake is nearly *isothermal*, there can be sufficient energy from the wind to mix the lake thoroughly. In the winter, *reverse stratification*, in which the deeper layer is *warmer* than the surface layer, may occur because water has a density maximum at 4°C (Fig. 2-8). As a lake cools in winter, the water temperature throughout becomes 4°C; then, as the surface water cools below 4°C, it becomes less dense than the deeper water. This stratification fosters the onset of ice, which prevents further wind mixing of the lake until the ice melts in the spring.

The effects of stratification on the lake environment are profound. During the warm season in a temperate zone lake, it is not unusual to have surface water temperatures between 15 and 25°C, while the lake remains only a few degrees above freezing near the bottom. Chemical reactions generally proceed more rapidly in the epilimnion due to warmer temperatures. Cold-water fish may be able to live in a particular lake even though its surface water temperatures are too high for their survival if the concentration of dissolved oxygen remains high enough in the hypolimnion (approximately 7 to 10 mg/liter, depending on species). Sufficiently high oxygen levels are commonly not maintained, however, especially in *eutrophic* (productive) lakes, in which large amounts of organic matter are produced in the epilimnion and settle into the hypolimnion. The isolation of the bottom waters from the atmosphere by stratification prevents the renewal of oxygen as it is consumed by organisms, and therefore the water may become *anoxic* or *anaerobic*. These two terms, anoxic and anaerobic, are often used interchangeably to imply an absence of molecular oxygen, O_2. Chemicals in an anoxic hypolimnion may undergo a set of chemical and biological transformations that are very different from those occurring in the epilimnion (see Section 2.4.3).

Vertical stratification due to water salinity differences also can occur in a lake; for example, in the winter, runoff water containing road salt can flow into a lake and travel along the bottom due to the greater density of the salty water, eventually accumulating in a low point of the lake and creating permanent stratification. Alternatively, natural processes, such as inflow of saline water from submerged springs, or accumulation of salts from chemical decomposition, may create permanent stratification. A lake in which this occurs is called a *meromictic* lake. Lack of mixing in meromictic lakes leads to a continuous depletion of oxygen in the bottom water (or *monimolimnion*).

2.2.3 ESTUARIES

Water Flow

Water flow in estuaries is more complicated than in rivers and in lakes; it is influenced by the inflow of freshwater from rivers and streams, by the tides of the ocean, and by the large salinity—and hence density—difference between fresh- and ocean water. The density difference tends to create a strong stratification, while the back-and-forth movement of water driven by the tides

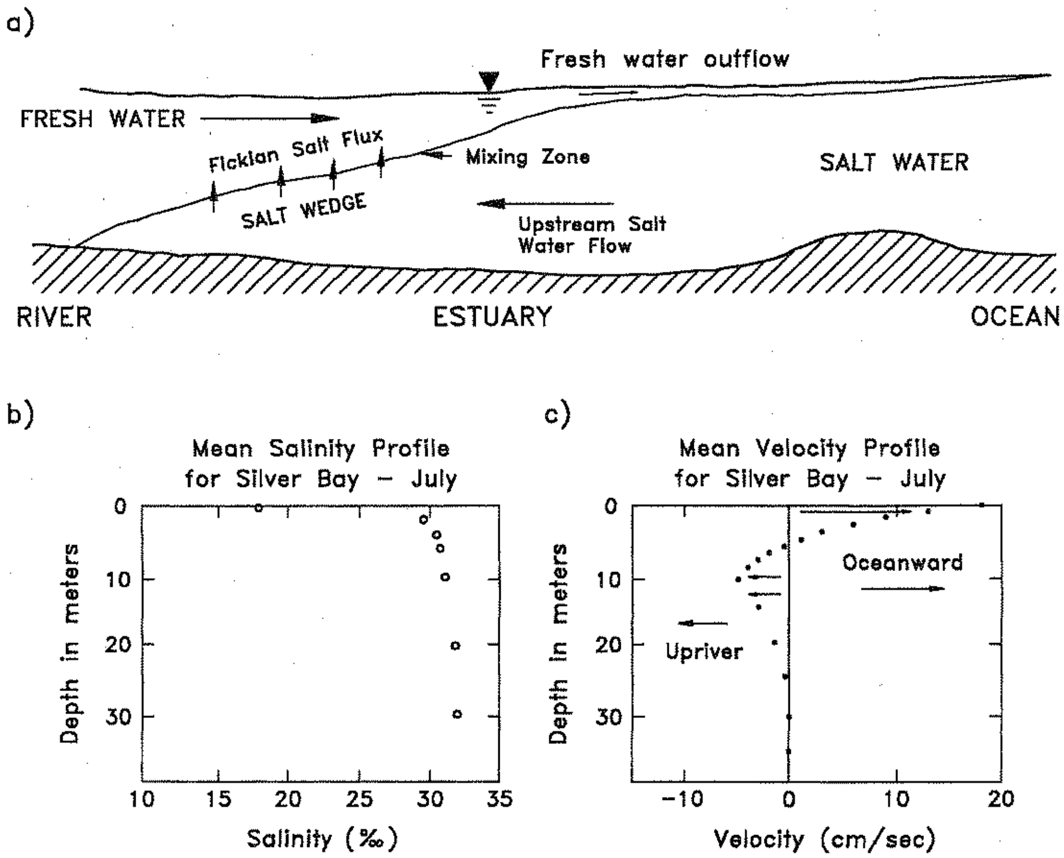

FIGURE 2-9 (a) An idealized estuary in cross section. In this well-stratified estuary, a distinct *salt wedge* extends upstream beneath fresher water at the surface. The freshwater/saltwater interface moves upriver at high tide and seaward at low tide. Data from Silver Bay, Alaska show (b) a steep average salinity gradient in parts per thousand (‰) and (c) the net upstream advection of saltwater at depth. More strongly mixed estuaries exhibit weaker vertical stratification [data are from Rattray (1967)].

enhances dispersion and mixing. In any estuary, *average* water movement is downstream (toward the ocean), driven by river flow; at any moment, however, the water can be flowing either way, depending on depth and whether the tide is rising or falling.

When a well-developed vertical stratification condition is established, an estuary is sometimes said to have a *salt wedge,* in reference to the underlying saltwater layer whose thickness tapers from the depth of the downstream saltwater body to zero in the upstream freshwater direction (Fig. 2-9). Interestingly, the average water velocity in a salt wedge is *upstream!* This can be shown by performing a mass balance for salt; upstream advection of salt by the saline layer must, on average, balance the outflow of salt in the freshwater layer. Salt is present in the freshwater layer due to upward diffusion from the salt wedge (i.e., Fickian transport).

Simple transport models have significant limitations in a complex estuarine setting; commonly, sophisticated numerical models are employed to predict transport in estuaries. In long, narrow estuaries, however, a simple one-dimensional model, such as is used in rivers, that incorporates a longitudinal dispersion coefficient and a time-averaged seaward water velocity can be useful. The results of such a model must be averaged over the tidal cycle; concentrations at each point in the estuary may be expected to vary significantly with the state of the tide. See Fischer *et al.* (1979) for a more complete discussion of transport in estuaries.

Stratification

Stratification in estuaries is in some respects similar to stratification in lakes, although in estuaries the density difference is primarily due to the difference in salinity between freshwater and ocean water, instead of being primarily due to temperature differences, as in most lakes. Freshwater has a density of approximately 1.00 g/cm^3, whereas ocean water has a density of approximately 1.03 g/cm^3 due to dissolved salts [primarily sodium (Na^+), chloride (Cl^-), calcium (Ca^{2+}), and sulfate (SO_4^{2-})]. This is a much larger density difference than that which occurs due to temperature differences in surface waters; hence, the stratification may be very strong. Whatever its cause, stratification always inhibits the vertical transfer of dissolved chemicals from layer to layer.

The salinity gradient in an estuary also has other effects on chemical fate and transport. As salinity increases in the region where freshwaters and saltwaters meet, particles brought in by the freshwaters tend to stick together (*flocculate*) and thus settle to the bottom more rapidly. The mechanism for increased flocculation is electrostatic. The rate of flocculation is given by the product of the frequency with which particles collide and the percentage of collisions that result in sticking. Particles of like charge repel each other, thereby decreasing the number of collisions that contribute to flocculation. A measure of the range at which repulsion occurs is the thickness of a "diffuse layer" of water, surrounding the particle, within which the charge of the particle is counterbalanced by a localized net excess of dissolved ions of opposite charge. As salinity increases, thereby increasing the ionic strength of the water, this diffuse layer decreases in thickness; this allows particles to approach each other more closely, resulting in more flocculation. Rising salinity also affects the activity of dissolved ionic chemical species due to the increasing ionic strength, thereby changing chemical equilibria in the water (see Section 1.6.5). Oxidation–reduction reactions are also affected because oxygen is less soluble in saline water (Table 2-4).

TABLE 2-4 Solubility of Oxygen (mg/liter) in Water Exposed to Water-Saturated Air at a Total Pressure of 760 mm Hg[a]

	Chloride concentration in water (mg/liter)				
Temperature (°C)	0	5,000	10,000	15,000	20,000
0	14.6	13.8	13.0	12.1	11.3
1	14.2	13.4	12.6	11.8	11.0
2	13.8	13.1	12.3	11.5	10.8
3	13.5	12.7	12.0	11.2	10.5
4	13.1	12.4	11.7	11.0	10.3
5	12.8	12.1	11.4	10.7	10.0
6	12.5	11.8	11.1	10.5	9.8
7	12.2	11.5	10.9	10.2	9.6
8	11.9	11.2	10.6	10.0	9.4
9	11.6	11.0	10.4	9.8	9.2
10	11.3	10.7	10.1	9.6	9.0
11	11.1	10.5	9.9	9.4	8.8
12	10.8	10.3	9.7	9.2	8.6
13	10.6	10.1	9.5	9.0	8.5
14	10.4	9.9	9.3	8.8	8.3
15	10.2	9.7	9.1	8.6	8.1
16	10.0	9.5	9.0	8.5	8.0
17	9.7	9.3	8.8	8.3	7.8
18	9.5	9.1	8.6	8.2	7.7
19	9.4	8.9	8.5	8.0	7.6
20	9.2	8.7	8.3	7.9	7.4
21	9.0	8.6	8.1	7.7	7.3
22	8.8	8.4	8.0	7.6	7.1
23	8.7	8.3	7.9	7.4	7.0
24	8.5	8.1	7.7	7.3	6.9
25	8.4	8.0	7.6	7.2	6.7

[a] American Public Health Association (1960).