CHAPTER 4

Meteorological and Other Non-Tidal Influences

4.1 Introduction

Because the tide usually dominates the spectrum of water level and current fluctuations along the ocean coasts, it is common to think of non-tidal fluctuations mostly in connection with inland waters. The tide in the deep ocean, however, can be quite insignificant, and, as shown in section 1.5 and Table 2, the tidal streams completely negligible from the standpoint of navigation. Wind-driven surface currents in the deep ocean, on the other hand, are of major importance to navigation. Water levels along ocean coasts are just as surely affected by atmospheric pressure and wind as are water levels along the shores of inland bodies of water. The range, however, is generally small compared to that of the tide on the coast, and the importance may not be fully realized until an extreme of the non-tidal fluctuations coincides with a corresponding extreme (high or low) of the tidal fluctuation. In using tidal predictions, such as those in the *Canadian Tide and Current Tables*, it should be borne in mind that they contain no allowance for non-tidal effects, other than for the average seasonal change in mean water level. The non-tidal influences are discussed below with reference both to ocean and inland waters.

4.2 Wind-driven currents

The major current systems of the ocean are driven by the wind stress acting on the surface. The direct effect of the wind stress is transmitted only to a limited depth by viscosity and turbulence, but the pressure gradients resulting from the induced surface slopes can set up deep flows in directions different from those of the surface flows. The main surface current systems of the Atlantic and Pacific oceans are in the form of large gyres that occupy most of the width of the ocean and are clockwise in the Northern Hemisphere and counter-clockwise in the Southern Hemisphere. The Coriolis acceleration is responsible for these circular patterns, deflecting both the winds and the currents driven by the winds. It may seem surprising that these "permanent" large-scale features of the ocean circulation could be the result of something as capricious as the wind. But, while the wind does vary from day to day with the passage of weather systems, it has a fairly consistent average pattern over much of the ocean, as witnessed by the Doldrums near the equator, the Trades in the tropics, and the Westerlies at mid-latitudes. Changes in the ocean current systems associated with seasonal changes in the average wind field are not well documented, except in certain areas such as the northern Indian ocean, where there is a marked difference between the current pattern during the southwest Monsoons of northern summer and that during the northeast Monsoons of northern winter.

Ekman, a Swedish mathematician and oceanographer, demonstrated that, in the absence of constraining boundaries, the surface current should flow in a direction 45° to the right of the wind stress in the Northern Hemisphere, and that over the whole water column there should be a net transport of water 90° to the right of the wind stress. Observations indicate that the wind-driven surface current flows at about 20° to 25° to the right of the wind as measured ten metres above the surface, and with a speed about two per cent of that of the wind. If, in the Northern Hemisphere, the wind blows parallel to a coastline on its right, the Ekman transport piles water against the coast until a surface slope is created to balance the Coriolis force; the current then flows parallel to the coast in the direction of the wind. If the coastline is to the left of the wind, the surface water is displaced away from the coast and deeper water rises to replace it. This "upwelling" is of biological importance in that it brings chemical nutrients back up into the euphotic zone (depth penetrated by sunlight), where they can be utilized in the growth of marine vegetation.

4.3 Wind set-up

The term *wind set-up* refers to the slope of the water surface in the direction of the wind stress. The slope perpendicular to a wind blowing along a coast, mentioned in the previous section, balances the Coriolis force on the along-shore current driven by the wind. It is an indirect effect of the wind, but is not usually thought of as wind set-up. When a wind commences to blow across the water surface, the wind stress is initially occupied in accelerating the water. When a steady state has been achieved, and the water is no longer accelerating, the balance of forces must be between the pressure gradient force due to the surface slope and the surface and bottom stress on the water (due to wind and bottom drag). Figure 35 illustrates the balance of forces in the direction of the wind stress for a wind blowing

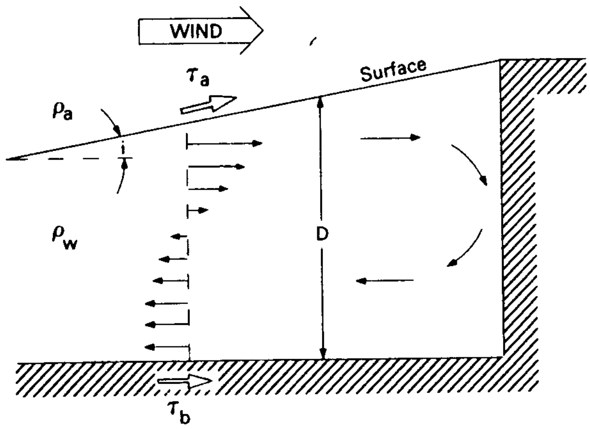

FIG. 35. Circulation of water and balance of forces associated with wind set-up.

toward shore, or along the axis of a lake or coastal embayment. The wind is assumed to have been blowing long enough for a steady state to have been reached, so the currents and the surface slope are not changing with time. The wind stress is just sufficient to maintain the surface current against the surface slope and the drag of the slower moving water beneath it. The pressure gradient caused by the surface slope is the same at all depths, and below a certain depth it drives a return flow. This flow is opposed by the bottom stress, or drag. The total drag force on a column of water of unit cross section and spanning the entire depth, D, is $\tau_a + \tau_b$, where τ_a is the surface wind stress and τ_b is the bottom stress. The horizontal pressure gradient is $\rho_w gi$, where ρ_w is the water density, g is gravity, and i is the inclination of the surface in the wind direction. The pressure gradient force on the total volume of the same column of water is therefore $\rho_w giD$. The bottom stress is usually considered small and proportional to the wind stress, and the balance of forces is approximately

$$(4.3.1)\ \tau_a = \rho_w giD$$

The wind stress is equal to the drag coefficient for air on water, C, times the density of air, ρ_a, times the square of the wind speed, W. We may thus write 4.3.1 as

$$(4.3.2)\ i = \frac{C\rho_a W^2}{\rho_w gD}$$

The drag coefficient is not a precise constant, but has a value of approximately 2×10^{-3}, ρ_a/ρ_w is 1.3×10^{-3}, and g is 10 m /s^2. Thus 4.3.2 becomes

$$(4.3.3)\ i = 2.6\times10^{-7}\left(\frac{W^2}{D}\right)$$

for W in metres per second and D in metres. Observations on lakes Ontario, Erie, and Huron, the Gulf of Bothnia and elsewhere indicate that the constant in 4.3.3 is too low, and should be between 4×10^{-7} and 5×10^{-7}. This is probably because of the neglect of the bottom stress in 4.3.1 and because of some funneling effects toward the ends of the lakes. Taking the larger experimental value for the constant, and expressing i as $\Delta h/L$, where Δh is the difference in water level over the length L, 4.3.3 gives

$$(4.3.4)\ \Delta h = 4.5\times10^{-7}\left(\frac{W^2L}{D}\right)$$

with all dimensions in metres and seconds.

We see from 4.3.4 that the difference in elevation between two ends of a lake, caused by a wind blowing along its length, is proportional to the square of the wind speed and the length, but inversely proportional to the depth. Wind set-up is thus of particular importance in shallow bodies of water with large horizontal extent. This is demonstrated by comparison of the values of Δh for Lakes Erie and Ontario given in Table 3 for various wind speeds along the lengths of the lakes. While the values in Table 3 have been derived from substitution in 4.3.4, examination of wind and water level data on these lakes confirms the relationship very closely. Because of the action of the Coriolis force, the axis of greatest slope would be slightly to the right (NH) of the wind direction. The line of flow of the surface current and the return deep current (Fig. 35) would also be oriented to the right of the wind.

4.4 Atmospheric pressure effect

The depression of the water surface under high atmospheric pressure, and its elevation under low atmospheric pressure, is frequently referred to as

TABLE 3. Samples of wind set-up on Lakes Erie and Ontario derived from equation 4.3.4

Wind speed	Δh(Erie)	Δh(Ontario)
2 m/s	0.04 m	0.00 m
5 m/s	0.26 m	0.02 m
10 m/s	1.03 m	0.07 m
15 m/s	2.32 m	0.16 m
20 m/s	4.13 m	0.29 m
25 m/s	6.45 m	0.45 m

Erie: length 390 km, mean of 1/depth = 1/17 m
Ontario: length 275 km, mean of 1/depth = 1/170 m

the "inverted barometer" effect. In a standard mercury-in-glass barometer, one atmosphere of pressure supports 0.76 m of mercury; if water were used instead of mercury in the barometer, the height of the column supported would be 10 metres. Since one atmosphere is approximately 100 kilopascals (kPa), we have the barometric equivalence of 10 cm of water and 1 kPa (or 1 millibar of pressure and 1 centimetre of water). Of course, if the water level is to rise in one place, it must fall in another; clearly, the level in a glass of water does not drop by 10 cm when the atmospheric pressure rises by 1 kPa. It is the slope of the water surface that adjusts to the atmospheric pressure gradient along the surface, so that in the absence of other forces, if the atmospheric pressure at *A* exceeds that at *B* by 1 kPa, the water level at *B* will exceed that at *A* by 10 cm. Stated another way, other forces again being absent, the water level at any location on a body of water differs from the mean surface level by an amount equivalent (but in the opposite sense) to the difference between the local and the average atmospheric pressure over the same body of water. The ocean is sufficiently large that it is fairly safe to assume that the average atmospheric pressure over its surface is constant, and that the inverted barometer effect is therefore fully experienced at each location. On lakes, however, a constant average pressure cannot be assumed, and water level differences from place to place must be treated instead of changes at one location only.

The change in water level caused by pressure change cannot easily be separated from that caused by wind set-up, because the winds are driven by the pressure gradients, and the two are closely correlated. It is usually best to assume that the pressure compensation is complete, and to credit the wind with the remaining surface slope. The justification for the assumption that the pressure compensation is complete is that the surface disturbance travels at the speed of a long wave, $(gD)^{1/2}$, which is usually fast enough to keep pace with moving weather systems. An interesting and useful result of the inverted barometer effect is observed in records from self-contained pressure gauges moored on the ocean bottom (section 6.7). Since these gauges are not compensated for atmospheric pressure, they record the sum of atmospheric and hydrostatic pressure. The compensation for changes in atmospheric pressure by changes in hydrostatic pressure, provided by the inverted barometer effect, is so nearly complete that most of the "noise" usually found in a tidal record disappears. Because there is no significant tidal signal in the water level fluctuations imposed by variations in atmospheric pressure, the loss of this part of the water level record simply leaves a cleaner record for tidal analysis. This is not, of course, a desirable feature if it is wished to record actual water levels for navigation or charting.

4.5 Storm surges

As the name suggests, storm surges are pronounced increases in water level associated with the passage of storms. Much of the increase is the direct result of wind set-up and the inverted barometer effect under the low pressure area near the centre of the storm. There is, however, another process by which the surge may become more exaggerated than would be anticipated from these two effects alone. As the storm depression travels over the water surface, a long surface wave travels along with it. If the storm path is such as to direct this wave up on shore, the wave may steepen and grow as a result of shoaling and funneling, as discussed for long waves in general in section 1.12. The term "negative surge" is sometimes used to describe a pronounced non-tidal decrease in water level. These could be associated with offshore winds and travelling high pressure systems, and are not usually as extreme as storm surges. Negative surges may, however, be of considerable concern to mariners, since they can create unusually shallow water if they occur near the low tide stage.

4.6 Seiches

A seiche is the free oscillation of the water in a closed or semi-enclosed basin at its natural period. They were discussed in section 1.6, and the formulae for the natural period of closed and open basins were given in equations 1.6.1 and 2. Seiches are frequently observed in harbours, lakes, bays, and in almost any distinct basin of moderate size. They may be caused by the passage of a pressure system over the basin or by the build-up and subsequent relaxation of a wind set-up in the basin. Following initiation of the seiche, the water sloshes back and forth until the oscillation is damped out by friction. Seiches are not apparent in the main ocean basins, probably because there is no force sufficiently coordinated over the ocean to set a seiche in motion. The tides are not seiches, being forced oscillations at tidal frequencies. If the natural period, or seiche period, is close to the period of one of the tidal species, the constituents of that species (diurnal or semidiurnal) will be amplified by resonance more than those of other species. The constituent closest to the seiche period will be amplified most of all, but the response is still a forced oscillation, whereas a seiche is a free oscillation.

A variety of seiche periods may appear in the same water level record because the main body of water may oscillate longitudinally or laterally at different periods, it may also oscillate both in the open and closed mode if the open end is somewhat restricted, and bays and harbours off the main body of water may oscillate locally at their particular seiche periods. Seiches generally have half-lives of only a few periods, but may be frequently regenerated. The largest amplitude seiches are usually found in shallow bodies of water of large horizontal extent, probably because the initiating wind set-up can be greater under these conditions. Table 4 lists a few seiche periods that have been observed and /or calculated for some Canadian waters. The list is by no means exhaustive, since seiches can be identified on almost any water level record; the entries in the table have been chosen simply to illustrate some of the principles mentioned above. The "typical large" ranges listed are not extremes, but are typical of perhaps the largest 10 or 20% of observed seiches. The last two entries in Table 4 are included not because of actual seiche activity in the Bay of Fundy, but because of interest in the part played by resonance in the large Bay of Fundy semidiurnal tides. For a long time study was concentrated on the resonant (or seiche) period of the Bay of Fundy alone, whereas it is now believed the important resonance is that of the tide with the oscillation of the combined Bay of Fundy – Gulf of Maine system.

TABLE 4. Sample seiche periods and typical large ranges.

Basin	Seiche type	Large range(m)	Period(h)
Lake Erie	closed	2	14
Lake Ontario	closed	0.2	5
Lake Huron	closed	—	6
North Channel of	closed	0.1	5
Lake Huron	open	0.1	10
Gore Bay (off North Channel)	open	0.1	0.2
Sydney Harbour,N.S.	open	0.3	2
Bay of Fundy	open	—	12
Bay of Fundy plus Gulf of Maine	open	—	13

4.7 Precipitation, evaporation, and runoff

The precipitation and evaporation considered here are those that occur at the water surface, not those that occur elsewhere in the drainage basin. The runoff is all the water that flows into the water system in question, and thus is the net result of precipitation, evaporation, and absorption of water over the land portion of the drainage basin. In the water budget of a system, precipitation is a positive term, evaporation a negative term, and runoff usually a positive term. In very arid regions, runoff could be negative by virtue of absorption of water into the parched soil along the shores; we are fortunate that in Canada this would be a rarity indeed. If the rate of input to the system from the sum of the three terms exceeds the rate of outflow at its mouth, the water levels within the system must rise, and, conversely, if the rate of outflow exceeds the rate of input, the water levels must fall. If there is no control on the outflow of a system, such as might come from dams, log jams or ice jams, the outflow would increase or decrease steadily with the rise or fall of the water level until an equilibrium was achieved between input and outflow. This is the basis for establishing "stage-discharge relations" from which the flow can be judged from the water level; they are valid only at locations below which there are no control structures or barriers.

There are seasonal variations in precipitation and evaporation that reflect in seasonal variations in water level and outflow of inland water systems, but the most dramatic changes are those associated with changes in runoff. Runoff reflects precipitation over the whole drainage basin, which may cover many times the area of the actual water surface. During a heavy sustained rainfall, only a portion of the water can be absorbed into the ground, and the runoff from a large drainage area can cause "flash flooding" of a water system. The Great Lakes system is unusual in that it has a small drainage area in relation to its large water surface, so no dramatic changes in level or outflow occur in that system. The water storage capacity of the land area of a drainage basin is greatly increased in winter, when much of the surface and ground water is locked up as ice, and the precipitation accumulates as snow cover. As the ice and snow melt in the spring, the runoff can increase rapidly, resulting in the spring "freshet" in streams and rivers, and in raised water levels throughout the system.

Sea level along open coasts is not noticeably affected by precipitation, evaporation, and runoff because their net average for the whole ocean is close enough to zero not to affect the elevation of such a large surface area. Water level records from a harbour at the mouth of a river may, however, reflect fluctuations in runoff. While the average water budget for the ocean is essentially zero, the local budgets are not, and water must be moved from place to place to minimize the occurrence of bumps and hollows on the surface. Significant currents are thus set up in the ocean to disperse water away from regions of high precipitation and/or runoff, and to divert water toward regions of excess evaporation (the ocean "deserts"). These currents do not flow directly from regions of budget surplus to regions of budget deficit, but are deflected, as are all currents, by the Coriolis force.

4.8 Effect of Coriolis force on currents

The Coriolis force was introduced and discussed in section 1.8. Its effect on tide propagation was considered in sections 1.8 and 1.10, and its effect on wind and wind-driven currents was mentioned in section 4.2. The fresh water runoff from land provides another example of the Coriolis force in action. Instead of continuing to flow directly away from the coast, this water is deflected to the right (NH) and forms a coastal current flowing along the coast to the right, for an observer facing seaward in the Northern Hemisphere (left in SH). In large lakes, bays and gulfs the runoff and the Coriolis force contribute to a cyclonic circulation, and around a large island they contribute to an anti-cyclonic circulation. The term "cyclonic" refers to rotation in the same sense as the earth's rotation on its axis. It is counter-clockwise when viewed from the Northern Hemisphere, and clockwise when viewed from the Southern Hemisphere. The term "anti-cyclonic" has just the opposite meaning. Use of these terms avoids continual reference to the reader's chosen hemisphere. The main wind-driven ocean current gyres are, in this terminology, anti-cyclonic. The Coriolis effect on currents is illustrated in surface current charts of the north Atlantic and Pacific (Fig. 36), the Gulf of St. Lawrence and East Coast (Fig. 37), and the Strait of Georgia (Fig. 38) between Vancouver Island and the B.C. mainland.

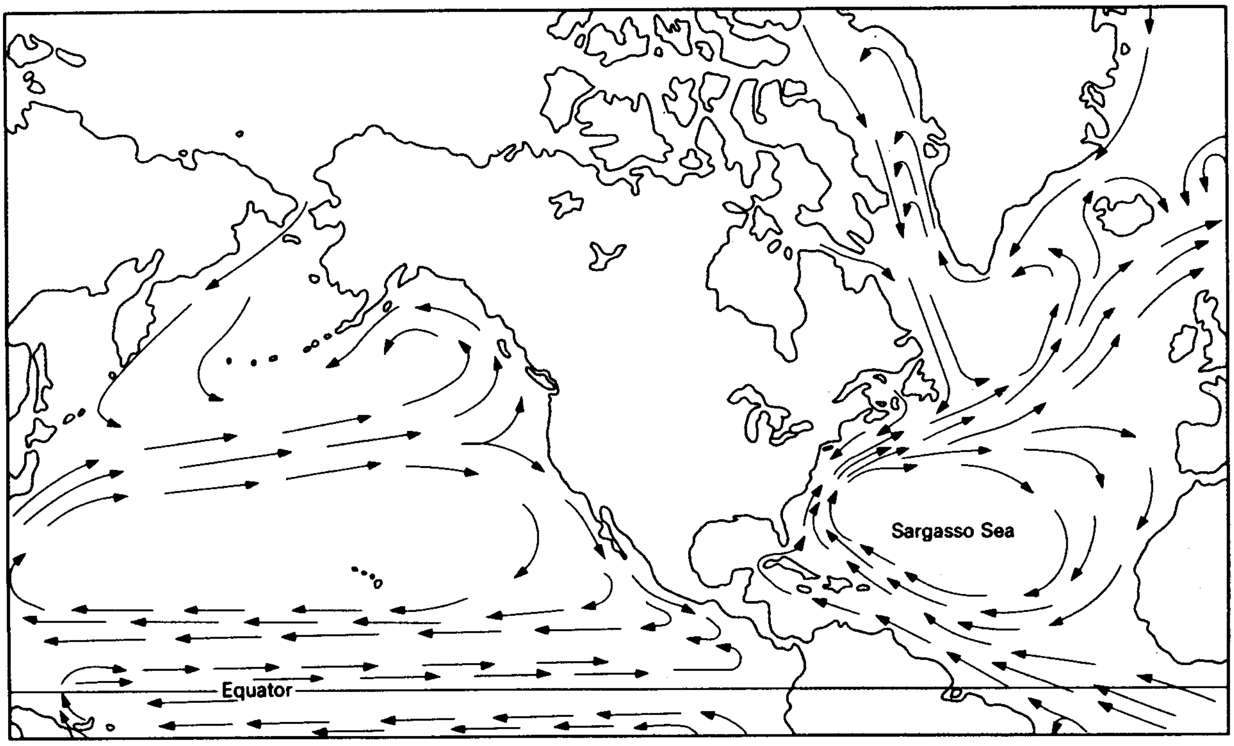

FIG. 36. Surface circulation in north Atlantic and Pacific Oceans.

FIG. 37. Surface circulation in Gulf of St. Lawrence and off East Coast.

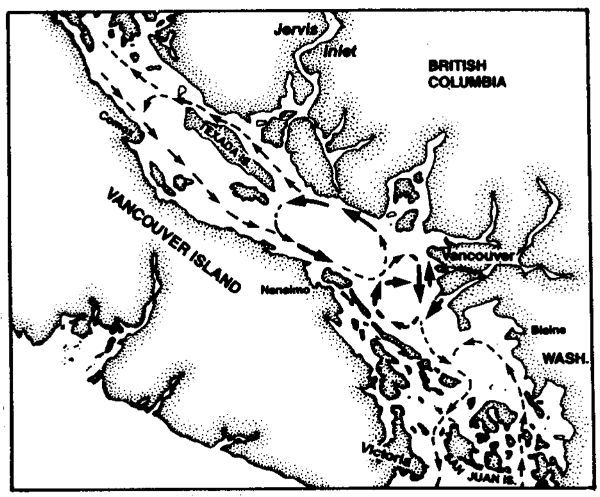

FIG. 38. Surface circulation in Strait of Georgia. (from fig. 10.23 of *Oceanography of the British Columbia Coast*, by R.E. Thomson).

It must be remembered that the Coriolis force does not generate currents, nor does it speed them up: its action is always perpendicular to the motion, and so can only change the direction of flow. The main anti-cyclonic gyres in both oceans south of latitude 45° N are driven by the anti-cyclonic wind stress and are reinforced in their anti-cyclonic pattern by the Coriolis force. The smaller cyclonic gyres north of latitude 45° N have this pattern partly because of the shape of the bathymetry and the coastline, and partly because the winds have a more cyclonic pattern at higher latitudes. The cyclonic circulations in the Gulf of St. Lawrence and Hudson Bay result from runoff and a slightly cyclonic average wind stress, assisted by Coriolis force, which likes to pile water up on the right-hand shore in the North. The circulation pattern in the Strait of Georgia fits the pattern of coastal runoff deflected to the right by the Coriolis force. In all of the above cases, the bathymetry of the basins also plays an important role in shaping the circulation patterns, but we shall not pursue that aspect here.

Inertial currents were the subject of section 1.9, but are mentioned here again because of their frequent appearance in current records from ocean moorings. They may be recognized in a record by their characteristic frequency, (12/sinφ) hours, where φ is the latitude. Inertial oscillations are not continuous in most records, but may appear and reappear several times, somewhat in the manner of a seiche in a water level record. The inertial signature is almost always present in current records from the deep ocean, but is rarely seen in records from shallow coastal regions. The sense of rotation is always anti-cyclonic around the inertial circle.

4.9 Estuarine circulation

An estuary is, for our purposes, any semi-enclosed body of water that has free access to the sea, a significant intrusion of sea water, and an inflow of fresh water. The mouths of rivers that flow into the sea are estuarine as far upstream as the limit of salt penetration, and most coastal harbours receive enough fresh water runoff to qualify as estuaries. The St. Lawrence system is estuarine from the limit of salt penetration at Quebec City through the Gulf to Cabot Strait and the Strait of Belle Isle. Estuarine circulation is a system of oppositely directed surface and deep currents driven by the mixing of the outflowing fresh water with the underlying salt water. Figure 39a illustrates in a simplified manner the principles of estuarine circulation; the fresh water is shown as if it entered only at the head of the estuary. The surface of the estuary slopes down toward the sea, and the fresher surface water flows seaward down this pressure gradient, becoming saltier as it mixes with the underlying water along its way. How much mixing takes place between the fresh and the salt water depends to a large extent upon the wind and the tidal action, but, qualitatively at least, the result is to tilt the isohalines (surfaces of equal salinity) down toward the head of the estuary as shown. The average density, which is roughly proportional to the salinity, is seen to be less for a column of water at the head of the estuary than for a column at the mouth. Because of this, the hydrostatic pressure increases more rapidly with depth at the mouth than at the head of the estuary, and the seaward-sloping pressure gradient near the surface may be replaced by a pressure gradient in the opposite direction at greater depths. This deep horizontal pressure gradient drives a return flow of saltier water into the estuary beneath the fresher outflow. If conditions remain unchanged for a sufficiently long time, a steady state would be reached in which as much salt is being transported into the estuary in the deep layer is being transported out in the surface layer, and the volume outflow of water exceeds the volume inflow by exactly the amount of the fresh water input. Without further consideration of the dynamics (i.e. the forces involved), we may find an interesting relation simply from the principles of continuity (i.e. the conservation of matter and the continuous nature of a liquid).

Consider a vertical cross-section of the estuary across the flow, and let S_o and S_i be the average salinities of the outflowing and inflowing water respectively, V_o and V_i be their volume transports, and R be the rate of volume input of fresh water. To conserve a steady state for the volume of water inside the estuary,

(4.9.1) $V_o - V_i = R$

and to conserve a steady state for the amount of salt inside,

(4.9.2) $S_o V_o = S_i V_i$

Solution of 4.9.1 and 2 for V_o and V_i gives

(4.9.3) $V_o = R\left(\frac{S_i}{S_i - S_o}\right)$

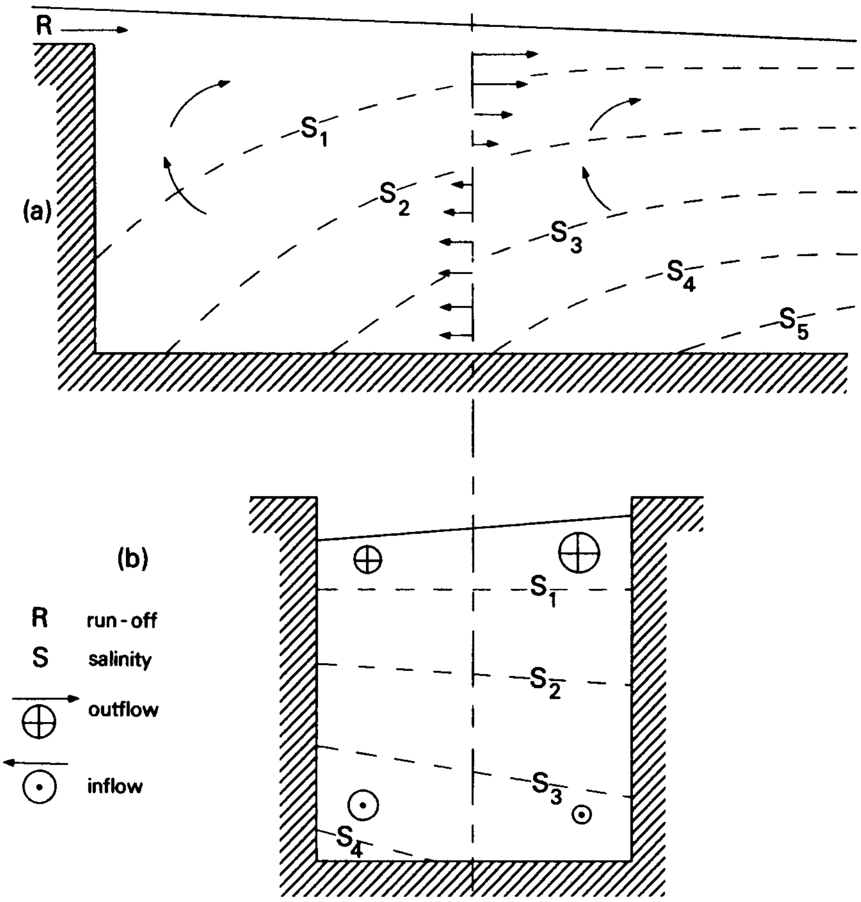

FIG. 39. Circulation and salinity patterns of estuarine circulation, (a) in a vertical section along the estuary and (b) in a vertical section across the estuary.

$$\text{and } V_i = R\left(\frac{S_o}{S_i - S_o}\right)$$

The volume transports in and out of the estuary depend critically on their salinity difference as well as on the fresh water input rate. The amount of mixing between the fresh and salt water along the estuary is therefore very important, since it determines the salinity difference. If we imagine the unrealistic situation in which there is no mixing, and the fresh water simply flows out over the undisturbed salt water beneath, S_o would be zero and 4.9.3 would give $V_o = R$ and $V_i = 0$. A more realistic example is the St. Lawrence estuary near Rimouski, Que., for which S_o is approximately 30‰ (parts per thousand), S_i is 34‰, and R is about 10 000 m^3/s. From these values, 4.9.3 gives V_o as 85 000 m^3/s and V_i as 75 000 m^3/s. Taking the width of the estuary at Rimouski as 45 km, the depth of the upper layer as 50 m, and that of the lower layer as 250 m, we may calculate the cross-sectional areas through which these volume transports are flowing, and so convert the transports into mean velocities in the two layers. This gives the mean outflow velocity in the upper layer as 0.04 m/s, and the mean inflow velocity in the lower layer as 0.01 m/s. These values agree well with observation, but they must be recognized as average values only. There is usually a strong vertical shear in the velocity, with the largest values near the surface, and both the outflow and inflow are usually stronger on their respective right-hand sides because of the Coriolis force.

The effect of earth rotation (i.e. Coriolis force) on a simple estuarine circulation is shown in Fig. 39b, a cross-sectional view of the estuary. The surface slopes up across the channel toward the right side of the outflow, and the isohalines slope down toward the same side. The strongest outlow is at the surface on the right side (NH) facing out. The strongest inflow is on its right side facing in, but is not always strongest at the bottom. The outflowing Gaspé Current (Fig. 37) is at least partly the result of intensification of the surface estuarine outflow along the Gaspé shore because of earth rotation.

The Mediterranean Sea is an example of a body of water in which the evaporation exceeds the sum of the precipitation and the runoff. Such a body of water does not qualify by our definition as an estuary, but the term "negative estuary" is sometimes applied to it, and the formulae in 4.9.3 may be used with the negative value of R to calculate what is now a surface volume transport inward and a deep volume transport outward. What happens physically in this case is that evaporation lowers the level of the surface, causing surface water from the ocean outside to flow inward down the slope. The evaporation also raises the salinity of the inside water, making the average density of a column of water inside greater than that of a corresponding column of water outside. Below a certain depth this reverses the direction of the pressure gradient, and drives a deep flow of high salinity water from the inland sea to the ocean. The effect of earth rotation is to tilt the surface up on the right side of the inflow and to tilt the isohalines in the opposite sense. Figure 40a and b illustrates the situation when R is negative. The picture is in every way similar to that in Fig. 39a and b, but with the directions of flow reversed. The warm and salty water that flows out from the Mediterranean through the Strait of Gibraltar can be detected at intermediate depths far out in the mid-Atlantic. In the above discussion, no mention has been made of temperature as a factor in determining the density of seawater. In the open

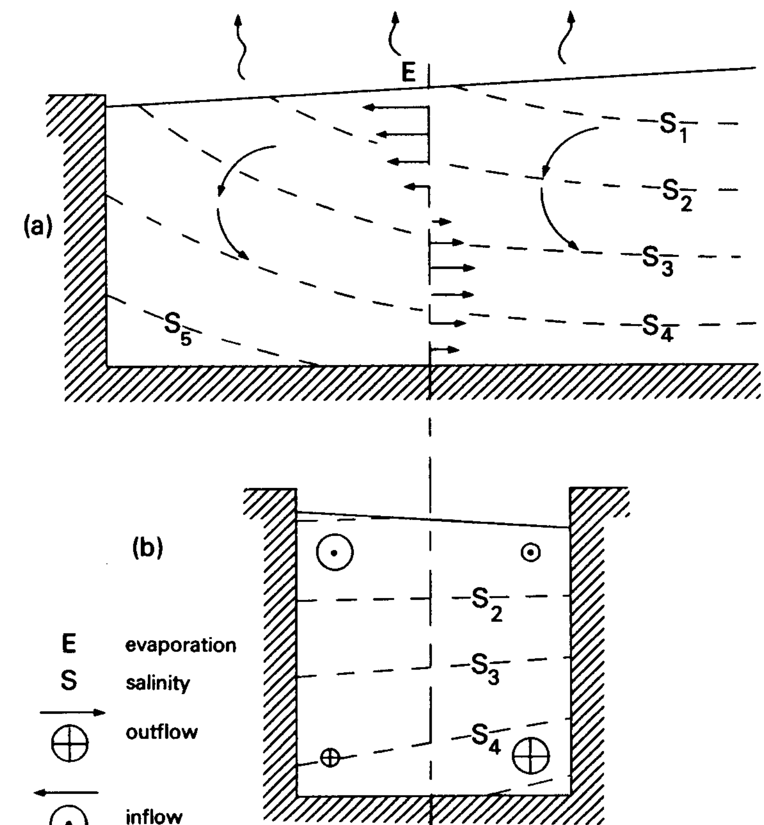

FIG. 40. Circulation and salinity patterns for a sea (e.g. Mediterranean) in which evaporation exceeds runoff plus precipitation, (a) in a vertical section along the axis and (b) in a vertical section across the axis.

ocean, where salinity differences are small, temperature is in fact the controlling factor for density, and colder water almost invariably underlies warmer water. In estuaries and other coastal regions, however, large salinity variations are common, and it is they that usually determine the density, with temperature inversions frequently occurring.

4.10 Melting and freezing

When seawater freezes, it is only the water that forms into ice crystals. The salt becomes trapped between the crystals in a concentrated brine that eventually leaches out, leaving mostly pure ice floating on the surface, surrounded by sea water of increased salinity and density. Since the ice displaces its own weight in this denser water, it does not displace as much volume as it occupied before freezing. Because of this, freezing has an effect similar to that of evaporation — it lowers the water level and increases the surface salinity and density. Surface water must therefore flow toward a region of freezing, while the cold salty water that is formed must sink and flow away from the region. In the polar regions, particularly in the Antarctic, freezing produces cold salty water that sinks and flows along the ocean bottom for thousands of kilometres. When sea ice melts, mostly fresh water is released, and this decreases the salinity and the density of the surrounding water. Melting thus has an effect similar to that of precipitation — it raises the water level and decreases the surface salinity and density. Surface water must therefore flow away from a region of melting ice. The speed of currents associated with freezing and melting in the ocean are never great.

4.11 Tsunamis

A *tsunami* is a disturbance of the water surface caused by a displacement of the sea-bed or an underwater landslide, usually triggered by an earthquake or an underwater volcanic eruption. The surface disturbance travels out from the centre of origin in much the same pattern as do the ripples from the spot where a pebble lands in a pond. In some directions the waves may almost immediately dissipate their energy against a nearby shore, while in other directions they may be free to travel for thousands of kilometres across the ocean as a train of several tens of long wave crests. Being long waves, they travel at the speed $(gD)^{1/2}$, giving them a speed of over 700 km/h (almost 400 knots) when travelling in a depth of 4 000 m. The period between crests may vary from a few minutes to the order of 1 h, so that in a depth of 4 000 m the distance between crests might range from less than a hundred to several hundred kilometres. The wave heights at sea are only the order of a metre, and over a wavelength of several hundred kilometres this does not constitute a significant distortion of the sea surface. When these waves arrive in shallow water, however, their energy is concentrated by shoaling and possibly funneling (section 1.12), causing them to steepen and rise to many metres in height. Not only are the tsunami waves high, but they are also massive when they arrive on shore, and are capable of tremendous destruction in populated areas. Because of the relative gentleness of tsunamis in deeper water, ships should always leave harbour and head for deep offshore safety when warned of an approaching tsunami. The origin of the word is, in fact, from the Japanese expression for "harbour wave." This name has been adopted to replace the popular expression "tidal wave," whose use is to be discouraged since there is nothing tidal in the origin of a tsunami. Another expression sometimes used for these waves is "seismic sea wave," suggesting the seismic, or earthquake, origin of most tsunamis.

A tsunami warning system for the Pacific has been established by the United States, with its headquarters in Honolulu, Hawaii. Other countries, including Canada, that border on the Pacific have since been recruited into the system. Canada's direct contribution consists of two automatic water level gauges programmed to recognize unusual water level changes that could indicate the passage of a tsunami, and to transmit this advice to Honolulu. The gauges are at Tofino on the west coast of Vancouver Island, and at Langara Island off the northwest tip of the Queen Charlotte Islands group. The tsunami warning centre at Honolulu receives immediate information from seismic recording stations around the Pacific of any earthquake that could possibly generate a tsunami; it calculates the epicentre and intensity of the quake and the arrival time of the as yet hypothetical tsunami at the water level sensing stations in the network; it initiates a "tsunami watch" at all water level stations in the path, for a generous time interval around the ETA of the hypothetical tsunami; and it issues tsunami warnings to the appropriate authorities in threatened locations if the water level interpretation indicates that a tsunami has indeed been generated.