Marine Ecology

Jeffrey S. Levinton

State University of New York
at Stony Brook

PRENTICE-HALL INC., Englewood Cliffs, New Jersey 07632

1 geography, physics, chemistry, and movement in the oceans

1-1 INTRODUCTION

The earth is truly a planet dominated by its oceans. Oceans cover approximately 71% of the earth's surface. About 80% of the surface of the Southern Hemisphere is covered by ocean whereas 61% of the Northern Hemisphere is oceanic. Most of the world's oceans are deep and 84% of the sea bottom lies at depths greater than 2000 m. The average depth of the oceans is 4000 m and some ocean trenches are as deep as 10,000 m. Inasmuch as our knowledge of marine organisms is, for all practical purposes, largely confined to those living at depths less than 100 m, our ignorance of most of the oceans is obvious.

The Pleistocene ice ages had dramatic effects on the marine biota. The periodic advance of continental glaciers during cold periods lowered sea level because of the transfer of some of the earth's water into ice. Fluctuations of sea level on the order of 100 m occurred (we are now in a period of relatively high sea level). Thus during the last glacial maximum 11×10^3 years ago the continental shelves of the world were mostly exposed to air. Semienclosed bays, such as Long Island Sound, were freshwater lakes or dry land. Retreats and advances of the ice also greatly changed the distribution of water masses and climatic belts in the ocean. Consequently, the recent evolutionary history of the marine biota has been in a framework of fluctuating planetary climatic change. This pattern is not peculiar to the Pleistocene but extends back through the history of the oceans and marine biotas. Glacial periods are known for the Ordovician and Permian; other climatic fluctuations have been documented as well. These patterns must have influenced the evolution of the marine biota.

In this chapter we introduce the ocean. We discuss its geography, topography, and water movements. Properties of seawater are described, with some consideration of the factors controlling these properties. We explain the effects of physical and chemical properties of seawater on marine organisms in Chapter 2.

1-2 GEOGRAPHY, TOPOGRAPHY, AND STRUCTURE

The major ocean areas are the Southern (Antarctic) Ocean, the Atlantic Ocean, the Pacific Ocean, the Indian Ocean, and the Arctic Ocean (Fig. 1-1). Smaller bodies of water known as seas are nearly enclosed by land or island chains and have local distinct oceanographic characteristics. Examples are the Gulf of Mexico, the Mediterranean Sea, the Baltic Sea, and the Japan Sea.

The Southern Ocean (Antarctic Ocean) is unique in having its major boundary with other oceans and in having a continuous water connection along lines of latitude around the earth. The northern surface boundary of the ocean is the subtropical convergence, where colder and more saline water descends northward below the surface. A general pattern of west winds in the 40 to 60° range of south latitude generates a surface current circling Antarctica (Fig. 1-1).

The Pacific is the largest ocean and is relatively little affected by the landmasses surrounding it. Island chains are most numerous in the Pacific and volcanic activity around the margins is pronounced. In contrast, the Atlantic is relatively narrow and is bordered by large marginal seas (the Gulf of Mexico, Mediterranean Sea, Baltic Sea, and North Sea). Its average depth is less than the Pacific Ocean. Furthermore, many of the great rivers of the world (the Mississippi River, Amazon River, Nile River, and Congo River) drain into the Atlantic system. The surface area of the Atlantic is only 1.6 times the surface of the land areas drained by the rivers flowing into it. The Indian Ocean is intermediate between the Atlantic and Pacific in depth and drainage.

Marginal seas often have unique oceanographic characteristics because of restricted circulation. Reduced mixing with the ocean permits local river drainage or the precipitation–evaporation balance to affect seawater properties. A shallow-water barrier, or sill, for instance, restricts circulation between the Atlantic and Mediterranean at the Strait of Gibraltar. Because of an excess of evaporation over precipitation and heat input, the Mediterranean is more saline and warmer than the adjacent Atlantic Ocean. The Baltic Sea, in contrast, has an excess of precipitation and river runoff over evaporation and averages only 100 m in depth. The salinity is low as a result. In other places (e.g., Black Sea, many Norwegian fjords), restricted circulation and consumption of oxygen by organisms result in deep waters devoid of oxygen.

The oceans share three main topographic features: (a) *continental margins,* composed of a shallow *continental shelf* and deepening *slope* complex, (b) *ocean basin floors,* and (c) *oceanic ridge systems*. The continental margin (Fig. 1-2) consists of the *shelf,* a low-sloping (1:500—about 0.1°) platform adjoining the continent and extending from a few miles to over a hundred miles from the shoreline. At the *shelf–slope break* (usually a depth of 100 to 200 m) the change in depth with distance from shore increases to 1:20 (about 2.9°), forming the *slope*. The slope is usually dissected by submarine canyons that act as channels for downslope transport of bottom material. Rapidly moving turbidity

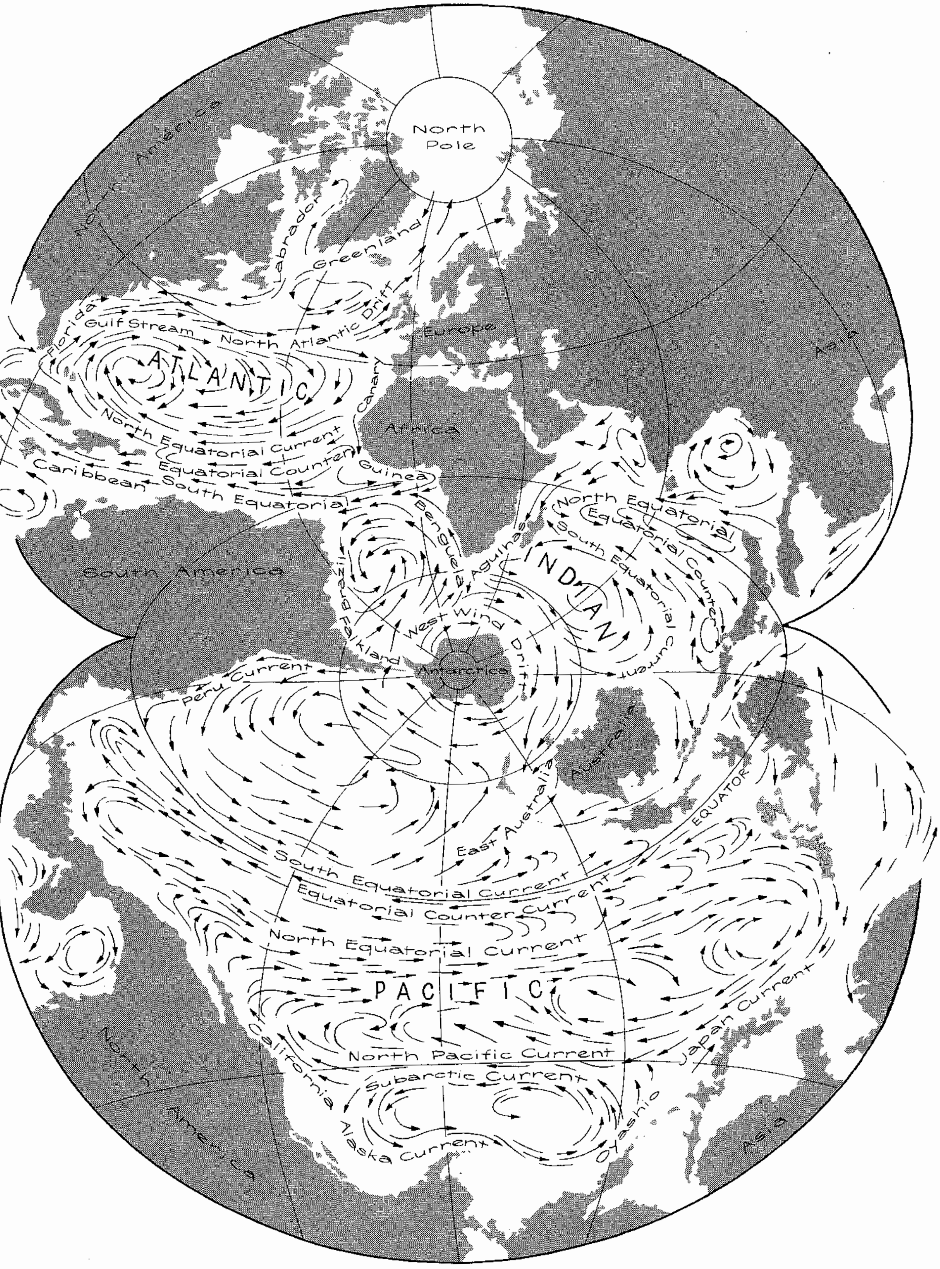

Figure 1-1 Surface currents of the world's oceans. (From "The Circulation of the Oceans" by Walter Munk, Copyright © 1955 by Scientific American, Inc. All rights reserved.)

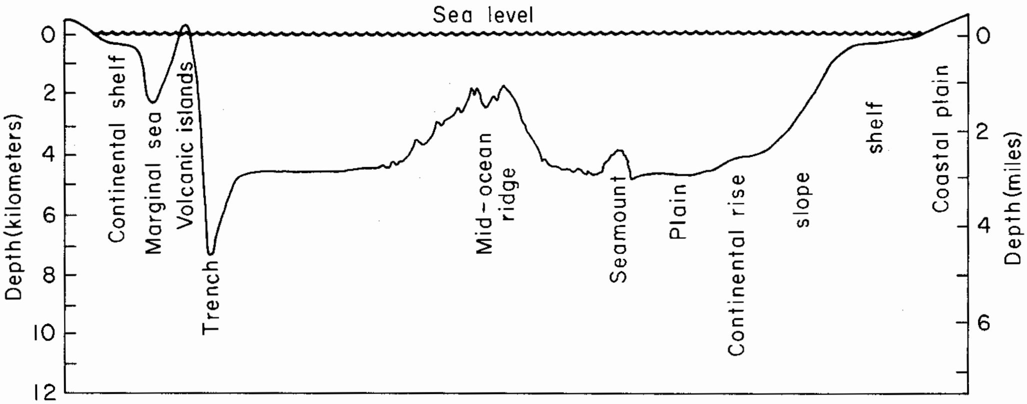

Figure 1-2 Two types of oceanic margins. From the right: shelf-slope rise. From the left: shelf-marginal sea-volcanic islands-slope-trench.

currents carry dense slurries of material. The slope descends to the deep-sea floor, but the topographic pattern at depth varies regionally. Two major sequences (Fig. 1-2) exist: (a) shelf–slope–rise and (b) shelf–slope–(marginal sea–volcanic islands)–trench. In the first case (e.g., the eastern North Atlantic), the foot of the slope merges with a deeper and gentler depositional feature known as a *rise,* which descends to the almost level *abyssal plain,* at an average depth of 4000 m. In the second case (e.g., the Aleutian Islands, Peruvian coast), the slope descends to a very deep (10,000 m) and narrow *trench,* parallel to shore. In the western Pacific the slope typically descends to a basin of about 2000-m depth, seaward of which is a chain of volcanic islands—island arcs—bordered by a still farther seaward trench. Seaward of these trenches is the abyssal plain, or ocean basin floor.

The *oceanic ridge systems* (Fig. 1-2) are a series of topographically high, linear ridges rising 2000 to 4000 m above the ocean basin floor. Usually they are submarine but often have emergent islands (e.g., Iceland, on the mid-Atlantic ridge). Ridges are volcanic in origin and are cut by transverse faults. Rift valleys, at the ridge system center, are parallel to the map trend of the ridge.

The deep ocean-basin floor consists of soft sediments derived from deposition of the following:

1. Mineralized skeletons of planktonic (freely floating) organisms, such as foraminifera, coccolithophorids, radiolarians, and diatoms derived principally from the upper 200 m of the water column.
2. Clay and other minerals that settle out of the water column to great depths.
3. Volcanic products consisting of basaltic rock and volcanic glass in various states, derived from oceanic volcanic islands.
4. Certain deposits actively forming through precipitation on the seabed, such as deposits of nodules of hydrous manganese and iron oxides. Shelf sediments derive

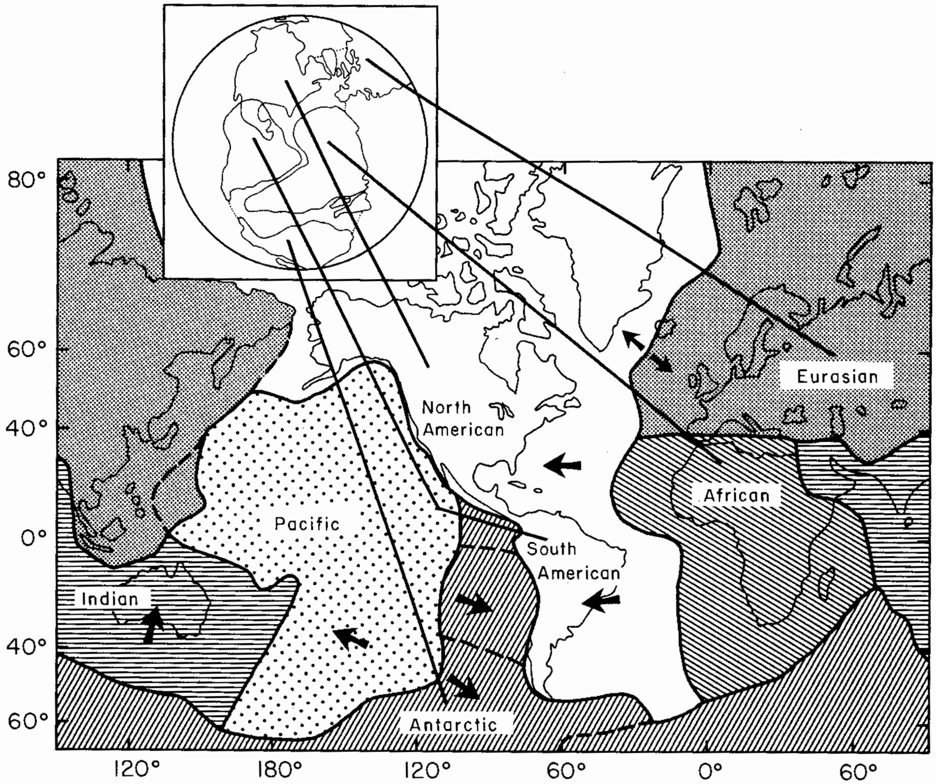

Figure 1-3 Sea-floor structure, showing plates, ridge systems, and directions of crustal movement. Inset shows arrangement of continents 225 million years before present.

from land erosion and from shelf sediments eroded subaerially during lower stands of sea level.

Beneath the soft-sediment mantle of the ocean basin floor is a crust made of (a) sediments in varying degrees of consolidation into sedimentary rock and (b) an underlying dense layer of volcanic rocks beneath the sedimentary cover. Several lines of evidence demonstrate that the ocean basin floors are moving horizontally (Fig. 1-3) a few centimeters per year. This phenomenon is known as sea-floor spreading. New crust is formed through volcanic activity at the oceanic ridges, carried as on a conveyor belt away from the ridges and consumed at trenches usually near ocean margins (Fig. 1-4). At the sites of crust destruction, crustal material is dragged downward and melted into the upper mantle below. The great depth of the trenches is a topographic reflection of the downward dragging of crust. The mechanism behind sea-floor spreading is probably a convective process involving a deep layer of the earth, the mantle.

The earth's crust consists of several major plates whose boundaries are trenches,

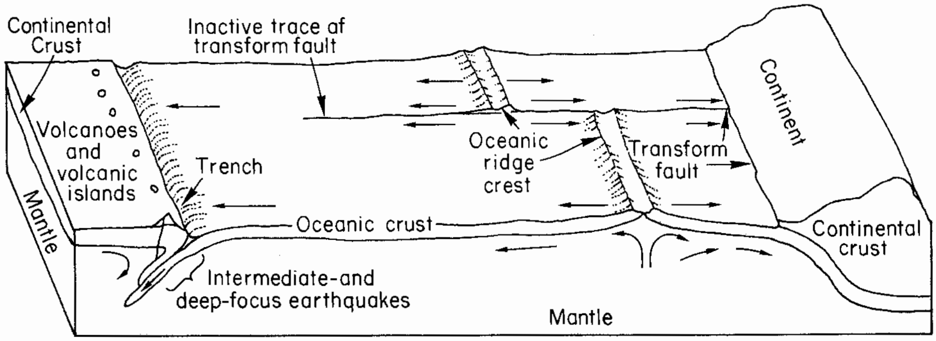

Figure 1-4 Schematic diagram of the oceanic crust, showing the formation of crust at ridges and downward transport-destruction at trenches.

ridges, or major breaks in the crust. Continents are enmeshed in plates and are conveyed along, via sea-floor spreading, in the plate in which they reside. So continents have drifted large distances over geologic time (Fig. 1-3).

Sea-floor spreading explains several well-known phenomena.

1. Volcanic rocks increase in age with distance to either side of a ridge system.
2. Oceanic sediments older than Cretaceous are relatively rare because they have been consumed.
3. Continents exhibit a jigsaw puzzle fit as if one large block had been fragmented (e.g., Africa and South America). This fit includes rock outcrop patterns and fossil distributions, as well as a general coastline match (see inset of Fig. 1-3).

The biological significance of these movements is based on these facts: (a) changes in the arrangement of the continents reorient current systems and water temperature regimes and (b) continental movement and changes in volcanic island systems probably influence biogeographic processes, such as dispersal and isolation of populations to specific ocean basins.

1-3 PHYSICAL AND CHEMICAL PROPERTIES OF SEAWATER

Water

Water molecules are asymmetric in charge distribution. Hydrogen bonds form between molecules, allowing a liquid state at atmospheric pressure and earth surface temperatures. Charge asymmetry allows water molecules to combine with other charged ions. Thus water is a good solvent for salts and biologically important molecules in cells.

Water has a high *specific heat;* that is, it takes 1 calorie to raise 1 gram of pure water by 1°C (at 15°C). Seawater of 35‰ at 17.5°C has a specific heat of 0.9. Thus it takes a great deal of heat to change the temperature of the ocean and the ocean can store and transfer a great deal of heat.

Salinity

Salinity (S) is the number of grams of dissolved salts in 1000 g of seawater (after all bromine has been replaced by chlorine, all carbonate converted to oxide, and all organic matter destroyed). It is expressed in *parts per thousand* (‰ or ppt) and ranges from 33 to 38 ‰ in the open ocean.

Because of the constancy of the *ratios* (see later) of major components, *chloride* has been used as an index of salinity. *Chlorinity* (Cl) is the number of grams of chloride ions in 1000 g of seawater. Chlorinity is measured by titration with silver nitrate (Knudsen method) and related to salinity through the following experimentally determined relationship:

$$S\ (‰) = 1.80655 \times \mathrm{Cl}\ (‰)$$

Seawater is an electrolyte and can therefore conduct an electric current. Seawater conductivity can thus be used as an index of salinity. At a given temperature electrical conductivity increases with salinity. Conductivity techniques of salinity estimation are far more accurate ($\pm 0.003\%$ for 32 to 38 ‰) than chlorinity measurements ($\pm 0.02\%$).

The refractive index of water is directly related to salinity. Modern refractometers are temperature compensated and provide a convenient alternative method for measuring salinity.

In the open ocean salinity is controlled by evaporation and sea-ice formation, which increase salinity, and by dilution processes, such as rainfall and river runoff. Latitudinal variation in precipitation and evaporation leads to two salinity maxima (Fig. 1-5) (ca. 35.5 ‰) at latitudes 30 N and 30 S. Slight excesses of precipitation relative to evaporation result in salinity minima at the equator (ca. 34.5 ‰) and at latitudes higher than 40 (ca. 34.0 to 34.5 ‰). Along coastlines more dramatic variation in salinity is conceivable because of the influence of river input.

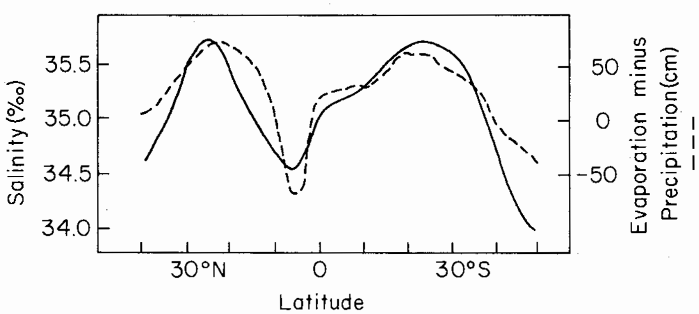

Figure 1-5 Latitudinal variation in surface salinity of the open oceans. Balance of evaporation and precipitation also shown. (After Sverdrup et al., 1942)

TABLE 1-1 MAJOR SEAWATER CONSTITUENTS

Constituent	Concentration, g/kg of 35 ‰ seawater	Constituent	Concentration, g/kg of 35 ‰ seawater
Cl^-	19.353	K^+	0.387
Na^+	10.76	HCO_3^-	0.142
SO_4^{2-}	2.712	Br^-	0.067
Mg^{-2}	1.294	Sr^{+2}	0.008
Ca^{+2}	0.413		

Salinity can increase in bodies of water when restricted circulation is accompanied by an excess of evaporation. Such conditions exist, for example, on the Bahama Bank west of Andros Island in the Bahamas, where the salinity is usually 40 ‰. Strong salinity changes also occur in tidal pools, where rainfall or evaporation can change seawater from 0 ‰ to concentrations at which various salts precipitate from solution.

Seawater is a complex solution containing nearly all known elements. The typical concentration of the main components of seawater is shown in Table 1-1. Chloride and sodium ions dominate. The relative abundances of all the components listed in Table 1-1 are nearly constant over almost all the world oceans despite overall changes in total concentration of solids. The close to constant ratios of concentrations of major ions is known as *Marcet's principle* or the *Forchhammer principle*. Such constancy is related to the residence time of an element in seawater or the mean length of time that a unit mass of a given constituent remains in the ocean. The respective residence times of most of the major ions in seawater are far greater than the time required to mix them evenly throughout the ocean.

Temperature

Temperature is a major factor regulating the distribution and abundance of marine organisms. The latitudinal thermal gradient is accompanied by major biogeographic changes in pelagic and bottom assemblages of organisms. At the lower extreme the freezing of seawater results in the formation of ice crystals that disrupt cells and terminate metabolic activity. At lethally high temperature physiological integration is impaired and enzymes are inactivated. Cytoplasm properties are altered and behavior is severely affected. Most marine organisms do not regulate their own body temperature (are poikilothermic). Temperature variation within lethal extremes thus has great effects on biochemical reactions and metabolism (discussed later).

Seawater has a much narrower range of temperature than air. Whereas air temperatures can range as low as −68.5°C (Siberia) to as high as +58°C (Libya), seawater ranges between −1.9°C (freezing point) to 40°C. The upper limit is relatively unusual and only obtained in water of a meter or less, as in sand flats of tropical Jamaica (Jackson, 1973). Open ocean seawater ranges from −1.9°C to about 27°C (Fig. 1-6).

The most significant factor affecting ocean water temperature is the latitudinal gradient of insolation or influx of solar energy. At low latitudes there is a net capture of

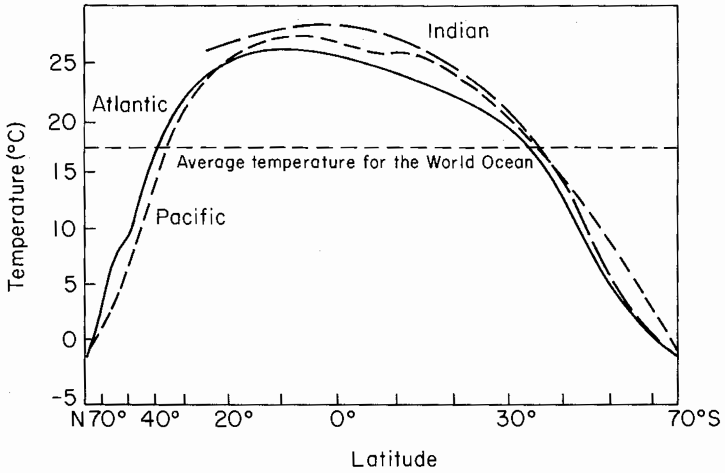

Figure 1-6 Latitudinal variation in sea-surface temperatures in the Atlantic, Pacific, and Indian oceans. (After Anikouchine and Sternberg, 1973)

heat from solar energy, but at high latitudes the earth loses heat. Other less important sources include (a) geothermal heating, (b) transformation of kinetic energy into heat (from water mass interactions), and (c) atmospheric processes, such as water vapor condensation and convection of heat from the atmosphere. Heat is lost to the atmosphere by (a) back radiation from the sea surface, (b) convection of heat to the atmosphere, and (c) evaporation. Most of the solar heat energy intercepted by the ocean is absorbed in very shallow water. So deep water is only about 2 to 4°C, even in the tropics. Latitudinal gradients in temperature are thus pronounced only in surface waters.

Surface water temperatures at the equator are seasonally stable as well as warm (Fig. 1-7). Seasonal fluctuations in the open ocean reach a maximum at intermediate latitudes (ca. 40 N and 40 S). At higher latitudes the temperature regime is again stable,

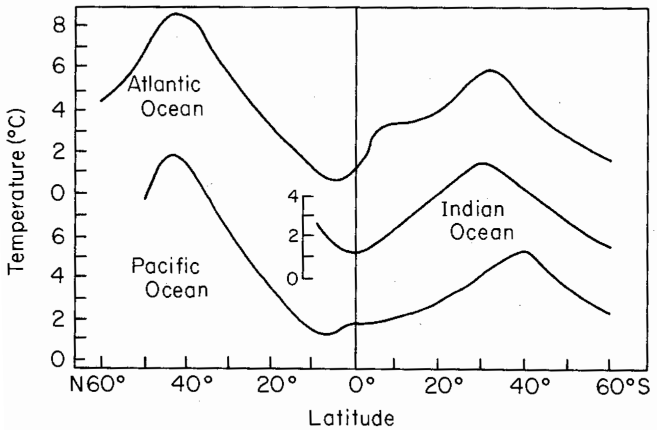

Figure 1-7 Annual range of sea-surface temperatures as a function of latitude. (After Sverdrup et al., 1942)

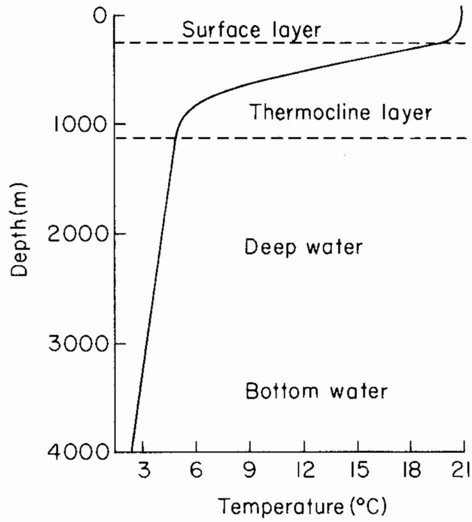

Figure 1-8 Temperature profile with depth in the open tropical ocean.

although cold. Seasonal temperature stability varies regionally as well; coasts whose climate is dominated by weather systems emanating from the continental interior (e.g., coast of New York) are less stable than coasts dominated by oceanic weather systems (e.g., West Coast of United States).

Because of solar heating, a stratified profile of temperature usually develops as in Fig. 1-8. The *surface layer* is warm. Due to wind-induced mixing, temperature of the surface layer is uniform with depth (isothermal). Another virtually isothermal *deep layer* is cooler. An intermediate *thermocline* is the depth range where temperature decreases from the surface layer to the deep layer. This overall vertical structure is generated by small eddy currents that transfer heat from the surface to deep water. The scale of this general structure varies with location. In the open tropical ocean the surface mixed layer is about 100 m deep, the thermocline extending to 1500 m. But this general vertical structure exists in small and relatively shallow bodies of water, such as Long Island Sound, New York (less than 50 m in depth), in spring and summer. Such vertical thermal structure induces a density gradient with cool dense water below and warm, less dense water above.

Density

Although density is usually expressed in grams per cubic centimeter (values ranging from 1.02400 to 1.03000 g cm^{-3} in the open ocean), the following convention is more convenient for oceanographically meaningful values:

$$\sigma_{S,T,P} = (\text{density} - 1) \times 10^3$$

Thus water with a density of 1.02400 would have a $\sigma_{S,T,P}$ of 24.0. This parameter, expressing density at a given salinity (S), temperature (T), and pressure (P) is usually abbreviated to σ_T ("sigma-tee"). Unless otherwise specified, σ_T refers to the density of water of a given salinity and temperature at atmospheric pressure (pressure increases density significantly).

Density increases with increasing salinity and decreasing temperature. At higher temperatures changes in temperature have more pronounced effects on density than at lower temperatures. The combined latitudinal variation of salinity and temperature results in a minimum surface density near the equator, with an increase northward and southward.

The depth zones of rapid temperature (thermocline) and salinity (halocline) change must also be regions of density change *(pycnocline)*. When water of low density resides above water of greater density, the water column is stable. In an estuary water of low salinity will float on top of denser, higher salinity water. In the presence of a thermocline, warmer and less dense water lies above colder and denser water. The stability imparted by this arrangement decreases the likelihood that wind can vertically mix the water column. Cold and saline surface waters, however, can be produced during sea-ice formation, as in the Arctic Sea and Antarctic Ocean. In this case, dense cold and saline water may sit on top of less dense water of lower salinity. This situation will be unstable, with sinking of the denser water.

Ice Formation

The formation of ice requires an orderly internal structure of water molecules. Ice crystals, however, are less dense (0.92 g/cm^3) than liquid distilled water (1.0 g/cm^3). Thus the formation of ice is accompanied by its flotation on liquid water. Although ice structure is loose, salts fit poorly in the crystalline matrix and are generally excluded during ice formation. Consequently, the formation of sea ice may increase the local seawater salinity. This situation occurs in the Weddell Sea in the Antarctic Ocean.

The presence of salts alters ice formation in two ways. First, the freezing point of seawater is lowered by increasing salinity. Secondly, water of salinity greater than about 15 ‰ will steadily increase in density as the temperature decreases toward the freezing temperature (-1.9°C). This situation is in contrast to freshwater, which has a density maximum at 4°C, with decreasing density as the temperature decreases toward the freezing point (0°C). Therefore seawater near the freezing point does not ascend in the water column due to density decrease, as occurs in freshwater lakes in winter.

pH of the Ocean

The acidity or pH of water is defined as the negative $\log_{10}$ of the activity of hydrogen ions. Because the activity of hydrogen ions in distilled water is 10^{-7}, the pH is 7. More acid waters have a lower pH whereas more basic waters have a pH greater than 7. Ocean water usually has a pH of about 8, making it a mildly basic solution. The pH of the ocean can be explained through chemical reactions involving dissolved carbon dioxide, carbonate and a few other weakly ionized chemical species, and crystalline calcium carbonate. Carbon dioxide is a source of carbon in terrestrial photosynthetic production of carbohydrates and is also the product of respiration. Table 1-2 shows the reactions in

TABLE 1-2 CARBONATE EQUILIBRIA

CO_2(dissolved) +	H_2O water	$\rightleftharpoons$	H_2CO_3 carbonic acid	
	H_2CO_3	$\rightleftharpoons$	H^+ hydrogen ion	+ HCO_3^- bicarbonate ion
	HCO_3^-	$\rightleftharpoons$	H^+	+ CO_3^{-2} carbonate ion
Ca^{+2} calcium ion +	CO_3^{-2}	$\rightleftharpoons$	$CaCO_3$(crystalline) calcium carbonate	

seawater. Reactions among these constituents result in alterations of the hydrogen ion concentration and hence the pH. If an equilibrium is being maintained between solution and precipitation of calcium carbonate, intermediate ions, and the pressure of carbon dioxide in the atmosphere, then the pH should be about 8—a close approximation to actual measurements. In cases where dissolved carbon dioxide is supplied in great quantities, as in the oxygen minimum zone, the pH may decrease to 7.5. In contrast, high local rates of photosynthesis in shallow water consume carbon dioxide and can raise the pH to 9.

1-4 CIRCULATION IN THE OCEAN

Coriolis Effect

All patterns of global water movement are affected by the earth's rotation. Note that the radius of a circular slice of the earth (perpendicular to the axis of rotation) at a given latitude decreases with increasing latitude. Because the earth rotates once a day on its axis, a particle at rest on the earth's surface at the equator must travel eastward more rapidly than a particle near the pole. A point on the equator has an eastward velocity of ca. 1700 km/hr while points at 30°N or 30°S have a velocity of ca. 1500 km/hr and points at 60° move at ca. 800 km/hr.

Consider a particle of water not attached to the earth, moving north from latitude 30°N. When it is at 30°N, it is moving eastward at a velocity of about 1500 km/hr. But as it moves northward, the earth beneath it is moving eastward at a slower velocity. Thus the particle will deflect toward the east, relative to the earth's surface, because of its initial higher eastward velocity. This process is called the *Coriolis effect* and is proportional to the sine of latitude. It is therefore zero in value at the equator and increases with increasing latitude toward the pole. It causes a deflection to the right for water traveling (in any direction) in the Northern Hemisphere and a deflection to the left for water traveling in the Southern Hemisphere (Fig. 1-9).

The Coriolis effect similarly deflects water moving under the force of the wind. Wind sets water in motion by dragging along the water surface, which is hydrodynamically rough because of ripples and other surface irregularities. Wind then causes water to move in sheets, which drag on layers of water below. The effect of the wind can thus be

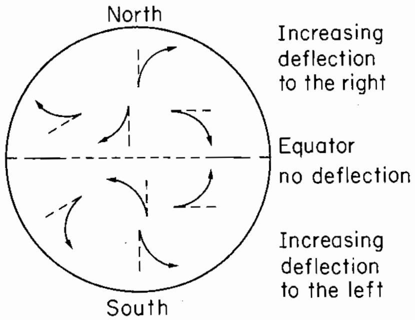

Figure 1-9 The Coriolis effect deflects moving objects, as shown by this diagram. Note that no deflection occurs at the equator.

transmitted to depths of 100 m, forming the surface layer discussed earlier. But the Coriolis effect deflects water to the right of the wind in the Northern Hemisphere (theoretically at an angle of 45°). Each layer of water is also deflected to the right of the direction of the immediately higher layer pulling on it. This results in a progressive deflection of a surface current produced by wind, with increasing depth, known as an *Ekman spiral*. Combining the movements of all layers usually results in a net 90° deflection of the surface layer from the direction of the wind (again, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere, characterized as *Ekman circulation*). This wind-induced surface transport is crucial to observed patterns of surface circulation in the ocean.

Because the surface waters are deflected approximately 90° to the right of the wind, less dense surface water tends to be pushed by the prevailing wind system in toward the center of clockwise (Northern Hemisphere) or counterclockwise (Southern Hemisphere) current systems known as *gyres* (Fig. 1-10). The deflection creates a higher sea surface at the center of, for example, the subtropical North Atlantic. This difference in elevation (less than 2 m) creates a gravitational pull. But as water flows down the inclined surface,

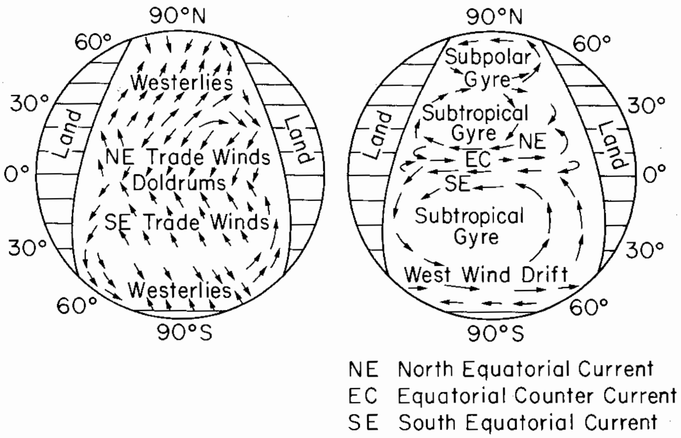

Figure 1-10 The relationship of the surface currents of the ocean to the planetary wind pattern. (After Fleming, 1957, courtesy The Geological Society of America)

it is deflected to the right until water flows in balance between gravity and the Coriolis effect. Such flow, known as *geostrophic flow,* is obtained imperfectly in the ocean.

Wind-Driven Surface Circulation

The latitudinal gradient in solar energy capture on the rotating earth drives the planetary wind system. At about 40°N and 40°S the *prevailing westerlies* help move surface water to the east while the *trade winds* move toward the west on either side of the equator.

Figure 1-10 shows the generalized surface circulation in the oceans and planetary wind patterns. The principal feature common to all ocean basins are gyres or large circular current patterns. The Atlantic and Pacific oceans have major gyres centered around the subtropical latitudes of 30°N and 30°S in each hemisphere. Because of the earth's rotation, gyres are displaced toward the west sides of oceans. Boundary currents on the west sides of oceans, such as the Gulf Stream, are stronger and narrower than diffuse eastern boundary currents. The warm north-flowing Gulf Stream transfers heat toward higher latitudes. Originating in the Gulf of Mexico, warm water travels northward along the coast of North America and then eastward toward the British Isles. Coastal waters of Ireland and the United Kingdom are warmed as a result.

Because surface water is transported away from the tropical western coasts of continents by wind, water rises from the bottom to replace it (Fig. 1-11). Such a process is known as *upwelling* and results in the transport of large amounts of nutrients from the bottom and great biological productivity in the surface waters.

Other conspicuous features of oceanic surface circulation are the east and west equatorial currents and the continuous water movement around Antarctica toward the east—the west wind drift.

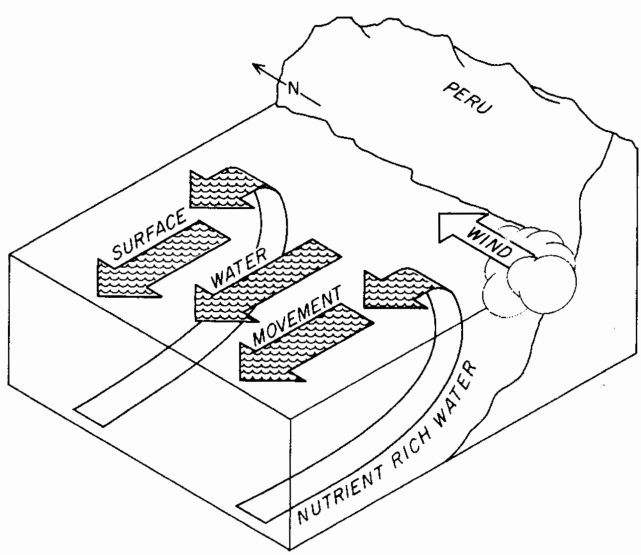

Figure 1-11 Vertical transport (upwelling) of coastal water as induced by winds combined with the Coriolis effect.

Thermohaline Circulation in the Deep Ocean

Although strong currents may exist locally in the deep ocean, deep oceanic circulation is dominated by movement of large *water masses* whose unique temperature and salinity characteristics are acquired during their origin at the surface at high latitudes. Figure 1-12 shows the origin and fate of major water masses of the Atlantic Ocean. The *Antarctic bottom water* originates at the surface in the Weddell Sea in Antarctica. Formation of nearly salt-free sea ice leaves a cold, saline residue water that sinks to great depths, mixing with other water along the way. This water moves northward along the sea bottom and can be traced to the Northern Hemisphere in the Atlantic. It also moves into the Pacific and Indian oceans. The *Antarctic intermediate water* forms at the surface near Antarctica and descends toward the north. It is, however, less dense than the North Atlantic deep water (NADW). NADW originates at about 60°N latitude, sinks and travels southward, overlying the Antarctic bottom water.

A water mass can be traced by its characteristic temperature and salinity acquired at the sea surface. Mixing between any two water masses may be traced as waters whose temperature–salinity characteristics are intermediate between two known water masses. Oxygen, which is consumed by organisms as the water moves, can also be used as a tracer to identify water masses. Finally, decay of radioactive carbon fixed by photosynthesis at the surface can also be used to measure rates of thermohaline convection. For example, C-14 dating has shown that it takes less than 600 years for the Pacific deep water to travel from 60 S to 30 N.

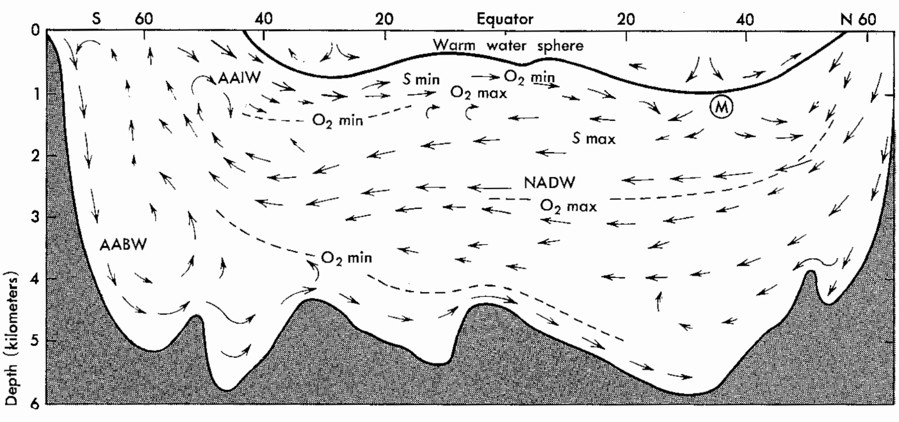

Figure 1-12 Thermohaline deep circulation of the Atlantic and Antarctic oceans, showing water masses identified by temperature and salinity. AABW = Antarctic Bottom Water; AAIW = Antarctic Intermediate Water; NADW = North Atlantic Deep Water. (From Gerhard Neumann, Willard J. Pierson, Jr., *Principles of Physical Oceanography,* © 1966, p. 466. Reprinted by permission of Prentice Hall, Inc.)

1-5 COASTAL PROCESSES

Waves

Ocean waves are generated by wind and deliver kinetic energy to the shoreline. Wave strength depends on the speed and duration of the wind and the distance of sea surface over which it moves. A series of waves passes a fixed point; peaks alternate with troughs. Figure 1-13 shows dimensions of waves. The time taken by two successive crests or troughs to pass a fixed point is the wave period T. L is the wavelength and H is the wave height. Wave speed V is calculated as

$$V = \frac{L}{T}$$

As a wave passes, water particles are set into vertical orbital motion, the diameter of the orbits diminishing to a zero radius deeper than $L/2$ (Fig. 1-13). Water movement in the direction of the wave is negligible relative to the lateral velocity of wave crests and troughs. As the wave approaches shore, vertical orbital movements of water particles strike the bottom when the depth is less than $L/2$. Here the wavelength (L) is shortened and the height (H) increases. Wave steepness (H/L) increases to a maximum of about 1/7 before the destabilization of the wave. At this point the wave "breaks" and expends energy through the formation of small sets of waves and turbulent water that wash on

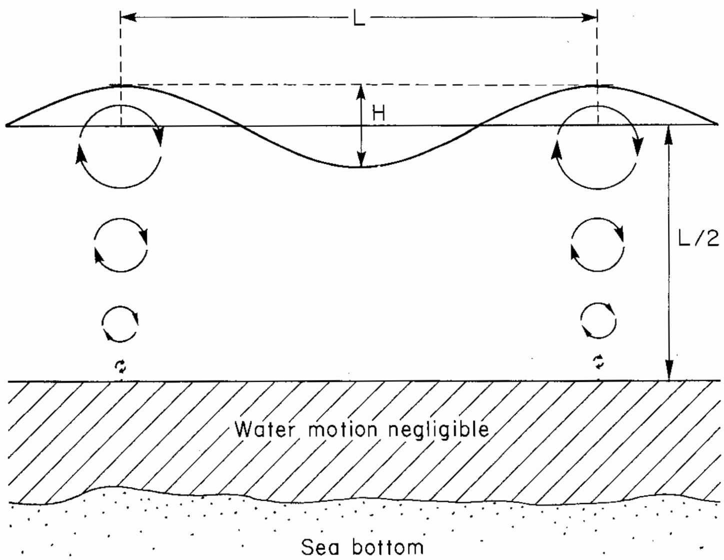

Figure 1-13 Dimensions of ocean waves. (After Shepard, 1963. From *Submarine Geology*, 3rd Edition by Francis P. Shepard. Copyright © 1948, 1963, 1973 by Francis P. Shepard. Reprinted by permission of Harper & Row, Publishers, Inc.)

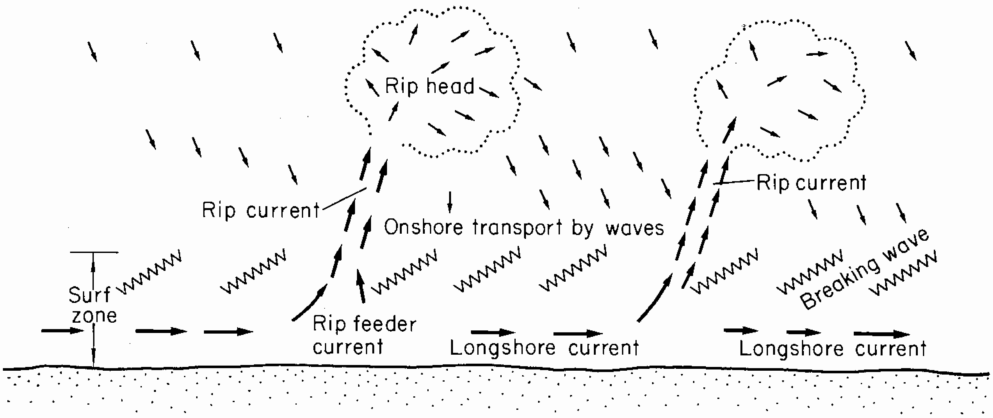

Figure 1-14 Water transport adjacent to an exposed beach. (After Shepard, 1963. From *Submarine Geology,* 2nd Edition by Francis P. Shepard. Copyright © 1949, 1963 by Francis P. Shepard. Reprinted by permission of Harper & Row, Publishers, Inc.)

the beach. This turbulence has enormous force. Organisms living on wave-swept coasts are sturdy and must have strong holdfasts.

As water approaches the shore, some of it is moved parallel to shore in the form of *longshore currents;* some moves rapidly offshore in concentrated *rip currents* (Fig. 1-14). Longshore currents are responsible for the transport of sediment parallel to the shore. The combined processes of breaking waves, wash on the beach, and longshore currents make sandy beach sediments unstable and an extremely rigorous environment for bottom-dwelling organisms.

As a wavefront approaches the shore, horizontal water velocity is less in shallower water relative to deeper water. Thus wavefronts refract as they approach irregular shores

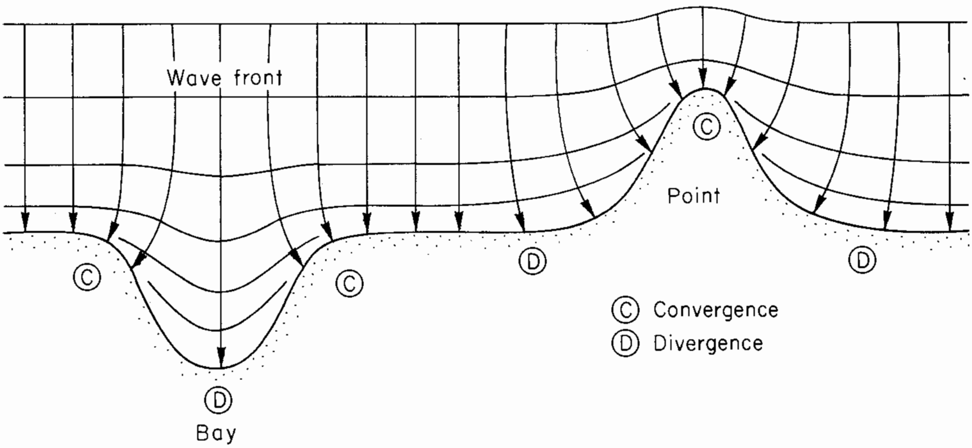

Figure 1-15 Wave refraction. Note that wave action is concentrated at headlands. (After Shepard, 1963)

(Fig. 1-15). Waves tend to concentrate on headlands that extend offshore as shallow protrusions, thereby hastening the smoothing of an irregular coastline. In contrast, when a wavefront enters a bay, the part of the front reaching the deep part of the bay travels faster than at either side of the bay. Thus wave energy diverges at depressions in the shoreline.

Although relatively rarer than typical wave action, large-scale storms, such as cyclones, hurricanes, and even high-force winds, can severely disrupt marine bottoms. Barrier islands are often breached in severe storms. Hurricanes can erode sediment in shallow water and topple large reef-building corals.

Tides

Although tidal motion is measurable in all parts of the ocean, to marine biologists it is only of great interest near the shore. Tidal rise and fall creates gradients of exposure to air and causes *tidal currents* that alter sediment distribution and local hydrodynamic forces.

Tidal movement is the result of the gravitational interaction of the earth, moon, and sun. The moon's proximity to the earth makes it the primary mass responsible for tidal motion. Gravitational attraction (force) between two bodies is proportional to the product of the masses of the two bodies, divided by the square of the distance between them. The attractive force between the earth and moon is balanced by a centrifugal force caused by the rotation of the two bodies about the earth–moon center of mass. Although in balance at the center of the earth, these forces are out of balance on the earth's surface to the extent that there is a net attraction to the moon on the side of the earth facing the moon and a net excess of centrifugal force on the side facing away from the moon (Fig. 1-16). Ideally this should result in a bulge of water (high tide) toward the moon and away from the moon (high tide) at any one time. Corresponding depressions (low tide) will exist on those parts of the earth where there is no net excess of gravitational pull relative to centrifugal force. Because the moon "passes over" any point on the earth's surface every 24 hours and 50 m, or one tidal day, ideally there should be two low and two high tides per day. Because the moon's position relative to the earth's equator shifts from 28.5°N to 28.5°S, the relative heights of high and low water differ geographically from gravitational forces alone.

During times when the sun, earth, and moon are in line (Fig. 1-16) the gravitational force exerted by the sun amplifies that of the moon. *Spring tides* occur at this time of maximal tidal range. When the sun, earth, and moon form a right angle, the two bodies cancel each other and *neap tides* occur, with minimal vertical range. The period between spring tides is 2 weeks.

The earth is not a sphere uniformly covered with water but a complex of oceans and smaller bodies of water interspersed by landmasses. The spatial arrangement of land and sea and the depth and size of basins affect the timing of tides and tidal heights. Moreover, the Coriolis effect also has a great influence on tidal heights and currents. Tidal currents, when affected by the earth's rotation, produce rotating systems around

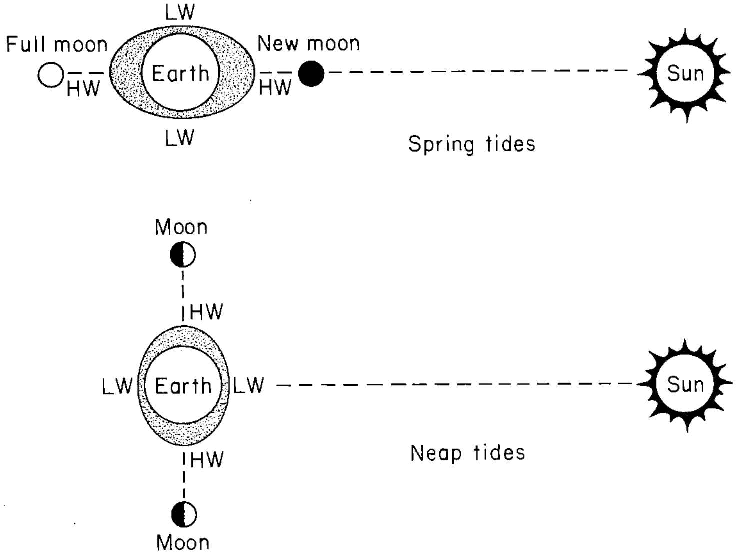

Figure 1-16 Action of tidal forces at different alignments of the sun and moon.

amphidromic points where no change in tidal height occurs over time. The North Sea has three such points.

The size of a given marine basin, relative to the time that a wave can travel from one end of the basin to the other, determines the response of the water in the basin to tide-generating forces. If travel time is short relative to the period of the tide-generating force, then the timing of the tide will be in equilibrium with that expected from an uninterrupted sphere of water. But if travel time is long relative to the tidal period, then the timing of the tide will be out of step with that expected for a uniform water-covered surface. Times of high tide may be quite different or even reversed from expectations based on an equilibrium prediction. One phenomenon commonly observed is the great difference in tidal timing of enclosed basins, such as bays, relative to the adjacent open ocean coast.

The various influences noted earlier, however, tend to cause deviations from the ideal *semidiurnal* tidal pattern. Nevertheless, an even pattern is approximately achieved in such areas as the East Coast of the United States and Southeast Asia. Yet in other parts of the world, such as the northeastern Pacific, the size of the basin seems to respond to a diurnal (solar) component of tidal action and there is only one pronounced low tide per day. In these systems, we speak of lower low water and higher low water. The condition of inequality of tidal highs and lows, where the semidiurnal component interacts with a diurnal component, is known as a *mixed tide*. Figure 1-17 shows some tidal patterns over several days under different tidal regimes.

Because of the various interference effects of landmasses, latitudinal effects, and basin size, some areas have almost no vertical tidal movement. The east coast of the

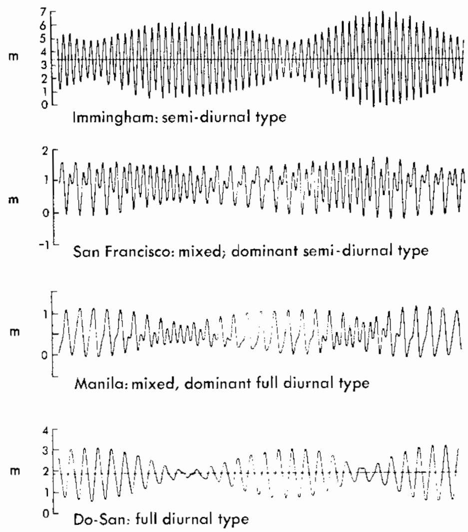

Figure 1-17 Tides from different locations of the world. (From Defant, 1960)

Jutland Peninsula of Denmark, for instance, has hardly any vertical water movement that can be related to tides. Uncoverings of sand flats are related to wind movements that push water toward or away from shore. The Caribbean coast of Panama also has a small tidal range. Consequently, when offshore winds uncover hard surface marine flats for days, mass mortality results from the heat and desiccation of the midday sun (Glynn, 1973b). So although regular tidal motion results in uncovering of marine organisms with concomitant exposure, it at least guarantees immersion some time in the day!

The rise and fall of the tide is accompanied by horizontal movements known as *tidal currents*. Near the coast, a current during the flood, or rise, of the tide is followed by a reverse current generated by the ebb, or fall, of the tide. Such currents can be strong and can transport pelagic larval stages, plankton, and oxygen throughout an embayment.

The vertical range of tides is also affected by the funnel-like nature of some enclosed basins. Usually these blind channels have a much greater vertical tidal range than adjacent open coasts. Dramatic examples include the Gulf of California, the Bristol Channel, United Kingdom, and the Bay of Fundy, Nova Scotia, where tidal range exceeds 10 m.

Ocean Fronts

The substrate facing the sea may be soft sediment or outcrops of rock under active erosion by wave and current action. In quiet water, soft sediment accumulates in *sand* or *mud flats* where sediment movement is minimal and the slope of the flat changes little throughout the year. Such flats can be hundreds to thousands of meters wide, as in the northern part of the Gulf of California, the Bay of Fundy, and the north coast of France. On more

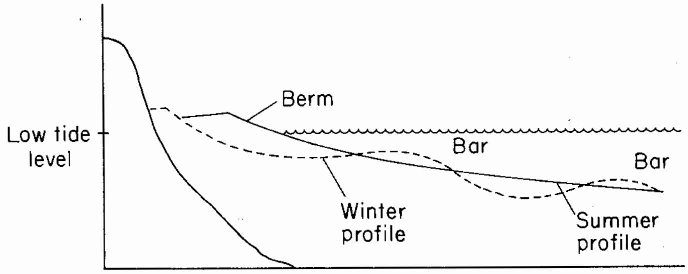

Figure 1-18 Seasonal differences in the profile of an exposed sandy shore.

exposed ocean fronts, *exposed sandy beaches* occur, where extensive wave action causes sediment transport and seasonal change of the beach profile.

Figure 1-18 shows some general features common to exposed beaches. A series of wind-blown dunes, sometimes stabilized by vegetation, lies behind the beach. A relatively horizontal shelf, the *berm,* extends to a break in slope. Descending from this break, the beach increases in slope, especially during winter storms, to the low tide mark. Seaward a complex of troughs and offshore sandbars develop. Some offshore bars become high enough to become islands. Beach complexes may protect lagoons or bodies of water landward of the beach. Longshore migration of sand is responsible for formation of barrier island complexes, such as the barrier beaches of Long Island, New York, which protect Great South Bay.

Rocky coasts develop where outcrops of rock occur in geologically youthful terrains at the sea–land boundary. The nature of such coasts depends on the local lithology (rock type) and wave action. Poorly lithified sandstones, for example, are often weathered and eroded into sand particles, leaving a sandy beach at the base of the rock outcrops (as at Santa Barbara, California). In contrast, highly cemented sedimentary rocks and crystalline rocks maintain their hard surface. Wave attack wears down cliffs at the shore, leaving a slope of debris subtidally beneath the cliff.

Estuaries

An estuary is a body of water nearly surrounded by land whose salinity is influenced by freshwater drainage. A gradient from open marine to freshwater, as in Chesapeake Bay, which extends from Virginia to Maryland, causes dramatic changes in the composition of marine biotas (see Chapter 17). Furthermore, rivers discharge large amounts of dissolved and particulate organic matter, providing nutrients that help sustain substantial estuarine fisheries, such as oysters in Chesapeake Bay.

Water movement in estuaries depends on the amount of river discharge, tidal action, and basin shape. All estuaries share a basic *estuarine flow,* where water of lower salinity moving seaward at the surface is replaced by more saline water moving up the estuary along the bottom. Usually some mixing by wind and tide occurs between the two layers. See Chapter 17 for details.

SUMMARY

1. Oceans cover most of the earth's surface. Ocean basins can be divided into continental margins, ocean basin floors, and oceanic ridge systems. Ridges are the sites of origin of volcanic oceanic rock; the ocean floor moves away from the ridge system in both directions. Such sea-floor spreading makes the earth's crust a dynamic layer; continents move along in crustal plates and change the shapes of ocean basins over millions of years.
2. Water's high specific heat, heat of vaporation, and transparency all permit large amounts of heat to be gathered, stored, and transferred between water masses and the atmosphere. Although the salt in seawater is a conservative property, increases are generated through evaporation and decreases stem from precipitation and river input. Salinity is very variable near coasts, where river input is significant.
3. Oceanic circulation is strongly influenced by the rotation of the earth. Surface circulation is driven by the planetary wind systems. Temperature and salinity variations at the surface influence the movement of water in the deep; cold and saline water moves from the surface at high latitudes into deeper waters and toward lower latitudes.
4. Ocean waves are generated by wind and deliver kinetic energy to the shoreline; orbital motion generated by waves can strongly affect coastal erosion and sediment transport. The influence of the moon and sun's gravitational attraction causes cyclic vertical tidal motion that is especially important at the coastline.
5. Estuaries are bodies of water in which a freshwater river mixes with the ocean. The lower density of freshwater, relative to seawater, generates a two-layered flow when tidal mixing is not too intense.