CHAPTER

7

The Ocean General Circulation and Climate

7.1 CAULDRON OF CLIMATE

The global ocean plays several critical roles in the physical climate system of Earth. These roles are related to key physical properties of the ocean, such as: it is wet, it has a low albedo, it has a large heat capacity, and it is fluid. Oceans provide a perfectly wet surface, which when unfrozen has a low albedo, and is therefore an excellent absorber of solar radiation. The oceans receive more than half of the energy entering the climate system, and their evaporative cooling balances much of the solar energy absorbed by the oceans, making them the primary source of water vapor and heat for the atmosphere. The world ocean is thus the boiler that drives the global hydrologic cycle. The world ocean also provides the bulk of thermal inertia of the climate system on time scales from weeks to centuries. The great capacity of oceans to store heat reduces the magnitude of the seasonal cycle in surface temperature by storing heat in summer and releasing it in winter. Because seawater is a fluid, currents in the ocean can move water over great distances as well as carry heat and other ocean properties from one geographic area to another. The equator-to-pole energy transport by the ocean is important in reducing the pole-to-equator temperature gradient. Horizontal and vertical transport of energy by the ocean can also alter the nature of regional climates by controlling the local sea surface temperature. The large heat content of the ocean and details of how heat is moved into the deep ocean are very important for the transient warming of the climate system, in response to changing climate forcing, such as the increase in greenhouse gases associated with human activities.

In addition to its direct physical effects on the climate system, the global ocean can affect the climate indirectly through chemical and biological processes. The ocean is a large reservoir of the chemical elements that form the atmosphere. Exchange of gases across the air–sea interface controls the concentration of trace chemical species containing oxygen, carbon, sulfur,

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and nitrogen, which are important in determining the radiative and chemical characteristics of the atmosphere. For example, the ocean controls the concentration of carbon dioxide in the atmosphere by exchanging gaseous carbon dioxide across the air–sea interface. Carbon dioxide is converted to organic solids in the ocean, and carbon is then stored by deposition of these solids onto the sea floor. Sulfur-bearing gases released from biological and chemical processes in the ocean enter the atmosphere where they are converted to aerosols that form the nuclei on which cloud droplets form. Evaporation of sea spray also forms salt particles that can form cloud condensation nuclei. The cloud condensation nuclei produced in the ocean can have a substantial influence on the energy balance of Earth, through their effect on the optical properties and extent of clouds.

7.2 PROPERTIES OF SEAWATER

Oceanic currents and the resulting heat transports are determined primarily by the physical properties of the ocean. To specify the physical state of seawater requires three variables: pressure, temperature, and salinity. Because water is slightly compressible, we define the potential temperature and potential density, which are the temperatures and densities at a reference pressure. As described in Chapter 1, *salinity* is the mass of dissolved salts in a kilogram of seawater, and is generally measured in parts per thousand, which we denote with the symbol ‰. The average salinity and temperature of the world ocean are approximately 34.7‰ and 3.6°C, respectively. The effect of ocean on atmospheric composition through biological and chemical processes depends on a more complex mix of physical, chemical, and biological properties. For example, oxygen and nutrient content of seawater are of critical importance for life in the sea. Trace amounts of key minerals may be very important for local biological productivity.

Variations of density on pressure surfaces are important for driving the circulation of the ocean, and depend on the temperature and salinity. Salt content increases the density of water, and seawater expands and becomes less dense as its temperature increases. The salinity of seawater ranges from about 25‰ to 40‰ and the temperature ranges from about –2°C to 30°C. Salinity and temperature variations have roughly equal importance for density variations in the ocean (Fig. 7.1). The density of seawater is almost linearly dependent on salinity. However, the dependence of density on temperature does not have a simple linear behavior. When the temperature of water approaches its freezing point, its density generally becomes less sensitive to temperature. For pure water, for example, the maximum density occurs at 4°C, and the water then expands slightly as it is cooled further. Therefore, fresh water lakes that are cooled from the top continue to overturn convectively until the entire water column reaches

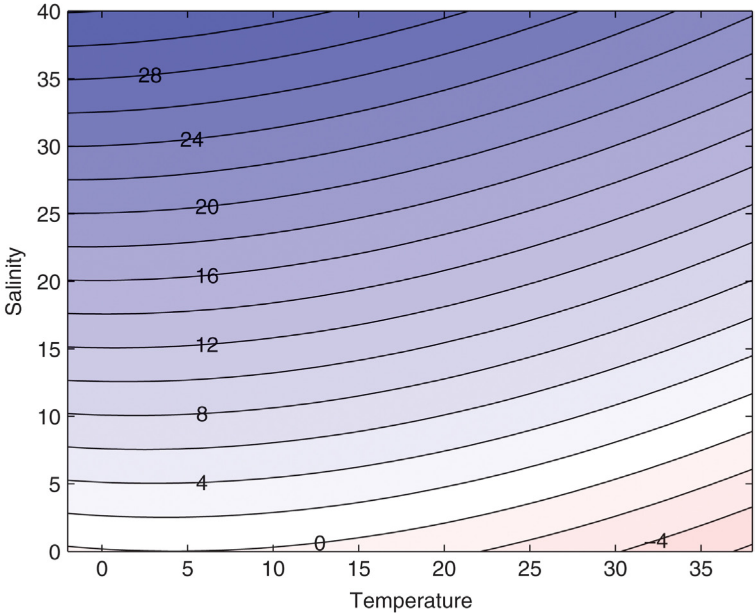

FIGURE 7.1 **Seawater density anomaly ($\rho_t - 1000\ kg\ m^{-3}$) at one atmosphere pressure as a function of temperature (°C) and salinity (‰).**

4°C, because water that is at 4°C will always be denser than warmer water. When the entire water column is cooled to 4°C, surface water that is cooled further will become less dense than the column and will "float" on the surface. When it reaches 0°C the surface water will freeze and form a layer of surface ice, which provides a floating layer of insulation between the cold air above and the warmer water below. If the lake is deep enough, the water near the bottom will remain at about 4°C, although the air temperature above the surface ice may fall to many degrees below zero. This fact allows fish in high-latitude or high-altitude lakes to survive the winter in the liquid water beneath the surface ice.

For seawater with salinity greater than 24.7‰, the density continues to increase with decreasing temperature until freezing occurs, although more slowly as the freezing point is approached. Therefore, if the salinity is initially uniform, the entire water column must be cooled to the freezing point before ice can form. Sea ice is able to form in high-latitude oceans because the salinity decreases significantly near the surface (Fig. 7.2). Lower salinities near the surface cause a decrease in density that offsets the increase in density associated with colder temperatures near the surface, allowing water near the surface to freeze while warmer water is present below. The low surface salinities result primarily from the excess of precipitation over evaporation in these latitudes. In the Arctic Ocean

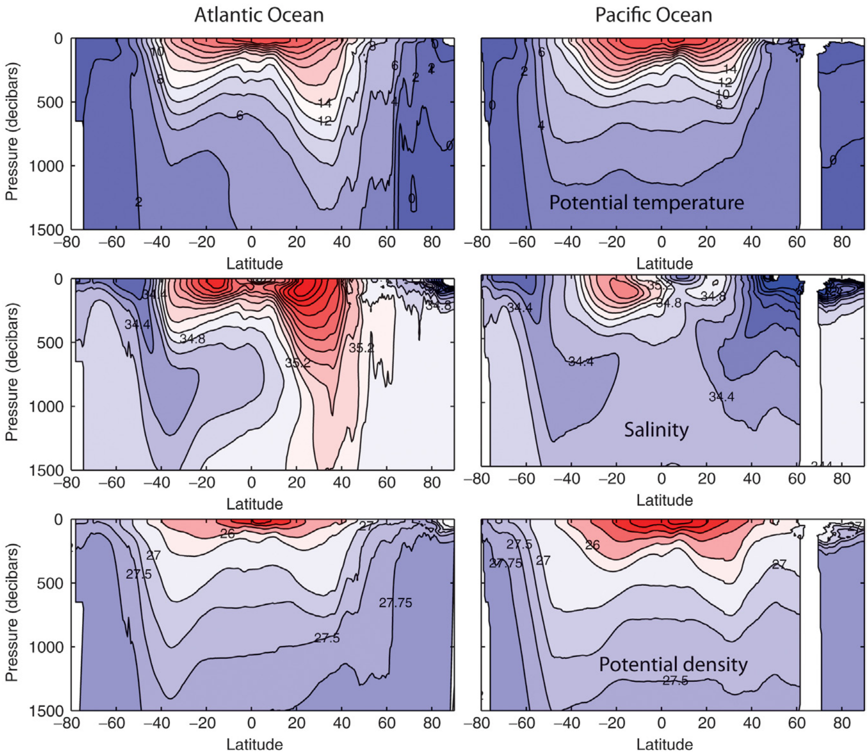

FIGURE 7.2 **Potential temperature (°C), salinity (‰) and potential density (kg m^{-3} – 1000) as functions of pressure (decibar ~1 m of depth) and latitude for the Atlantic (left) and Pacific (right) oceans.** *Data are from MIMOC.*

the supply of freshwater from rivers flowing from the surrounding continents contributes importantly to low surface salinities and therefore to the stable density gradient. The Arctic Ocean has a strong *halocline*, with relatively fresh water near the surface rapidly transitioning to more salty water beneath. Because salinity increases with depth and also increases density, the surface is able to form ice without bringing the temperature of the entire water column to the freezing point. It has been hypothesized that, if the flow of certain key rivers were diverted from the Arctic Ocean to supply irrigation water to continental interiors farther south, the heat balance of the Arctic could be severely distorted, because the normal configuration of a thin layer of surface ice on a mostly unfrozen Arctic Ocean may no longer be stable. Increased salinity of surface waters in the Arctic might lead to either complete removal of most Arctic sea ice or complete freezing of the Arctic Ocean from surface to bottom.

Only about the first kilometer of ocean between 50°N and 50°S is warmer than 5°C (Fig. 7.2), so that much of the mass of the ocean is between

–2°C and 5°C. The thermal structure of the ocean at most locations can be divided into three vertical sections. The top 20–200 m of water in contact with the atmosphere usually has an almost uniform temperature, which is maintained by rapid mixing through mechanical stirring and thermal overturning. This layer is called the surface *mixed layer* of the ocean. Below the mixed layer, the temperature decreases relatively quickly with depth down to about 1000 m (Figs 1.11 and 7.2). This layer of rapid temperature change, called the *permanent thermocline*, persists in all seasons. It is believed that the permanent thermocline is maintained by heating from above, balanced by a slow upward movement of colder water from below, and by very weak mixing at depth. The cold water in the deep abyss of the oceans is produced at the surface in a few regions of the polar ocean. At the base of the permanent thermocline, the typical temperature is about 5°C, and below this the temperature decreases more slowly with depth, reaching a temperature of about 2°C in the deepest layers of the ocean. The physical properties of the deep ocean show little spatial variability, so that temperature, salinity, and density are almost uniform (Fig. 7.2).

Water is almost incompressible, so the density of seawater is always very close to 1000 kg m^{-3}, even near the bottom of the ocean where the pressure may be several thousand times the surface air pressure. Density of seawater is usually reported as a deviation from 1000 kg m^{-3}, $\rho - 1000$. *Potential density*, ρ_t, is the density that seawater with a particular salinity and temperature would have at zero water pressure, or the density at surface air pressure. Potential density increases most rapidly with depth in the first several 100 m of the tropical and mid-latitude ocean (bottom row in Fig. 7.2). This rapid increase of density with depth is supported by the absorption of solar radiation near the surface, which sustains the warm temperatures there. The strong density stratification in the upper ocean inhibits vertical motion and turbulent exchanges, so that the deep ocean is somewhat isolated from surface influences in those regions where this density stratification is present. The strong density stratification is reduced in high latitudes, where in some locations the potential density at the surface comes much closer to the densities prevailing in the deep ocean.

The distribution of potential density suggests that water occupying the bulk of the deep ocean came from the Polar Regions, where at certain locations and seasons, surface water becomes dense enough to sink to a great depth. The distributions of other tracers also suggest that slow downward motion of water in high latitudes extends downward and equatorward into the deep ocean, as will be discussed in Section 7.6. Figure 7.2 shows that the potential density in the North Atlantic is greater than that in the North Pacific, especially near the surface where deep water is formed. This is because the Atlantic Ocean is more saline than the Pacific. So deep water can be formed in the Southern Ocean poleward of 50°S, and in the North Atlantic, but not in the North Pacific at the present time. At around 50°N in the Pacific and 50°S, fresh water is transported downward and spreads equatorward below the saline waters of the subtropics. In the Atlantic Ocean, relatively salty water persists below the surface layer of less saline water.

7.3 THE MIXED LAYER

The primary heat source for the ocean is solar radiation entering through the top surface. Almost all of the solar energy flux into the ocean is absorbed in the top 100 m. Infrared and near-infrared radiations are absorbed in the top centimeter, but blue and green visible radiation can penetrate to more than 100 m if the water is especially clear. The depth to which visible radiation penetrates the ocean depends on the amount and optical properties of suspended organic matter in the water, which vary greatly with location, depending on the currents and local biological productivity. The principal component of suspended matter in surface water is *plankton*, which are plants and animals that drift in the near-surface waters of the ocean. The relative abundance of photosynthetic plankton can be estimated from the color of the ocean whose greenness indicates the abundance of chlorophyll (Fig. 7.3). Chlorophyll is most abundant where ocean currents or mixing bring nutrients to the surface where they can be consumed by plankton.

The solar flux and heating rate in the ocean are greatest at the surface and decrease exponentially with depth, in accord with the Lambert–Bouguer–Beer Law, as described in Chapter 3. Under average conditions, the solar flux and heating rate are reduced to half of their surface value by

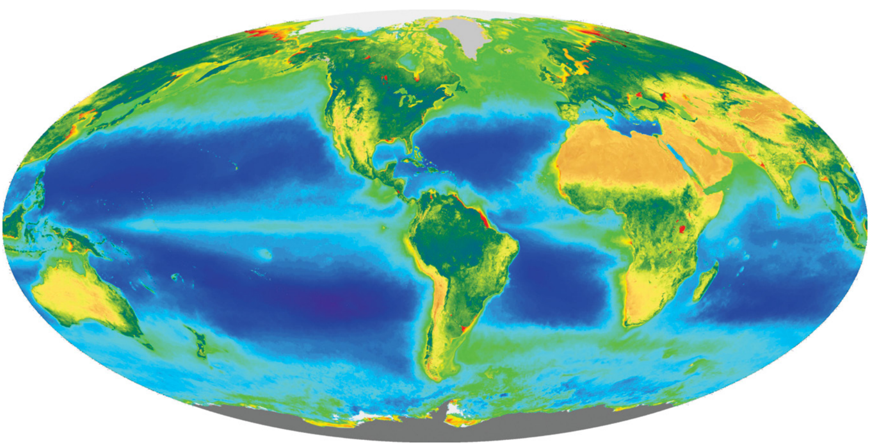

FIGURE 7.3 **The global biosphere – ocean chlorophyll concentration and land vegetation index derived from seaWiFS satellite data for September 1997–August 2000.** *NASA image: Mollweide projection.*

a depth of about 1 m, but significant heating can still be present at more than 100 m below the surface. Since the solar heating is deposited over a depth of several tens of meters in the upper layers of the ocean, and cooling by evaporation and sensible heat transfer to the atmosphere occurs at the surface, an upward flux of energy in the upper ocean is required to maintain an energy balance between surface loss terms and subsurface heating. Molecular diffusion is an important heat transport mechanism only in the top centimeter of the ocean. Elsewhere the heat flux is carried by turbulent mixing, convective overturning, and mean vertical motion, which is called *upwelling* or *downwelling* in the ocean. Turbulent mixing in the surface layer of the ocean is greatly aided by the supply of mechanical energy by the winds and their interaction with waves on the surface of the water. In the mixed layer of the ocean, heat transport by convection and turbulent mixing is so efficient that the temperature, salinity, and other properties of the seawater are almost independent of depth.

A schematic diagram showing the processes important in the oceanic mixed layer is presented in Fig. 7.4. The depth of the mixed layer depends

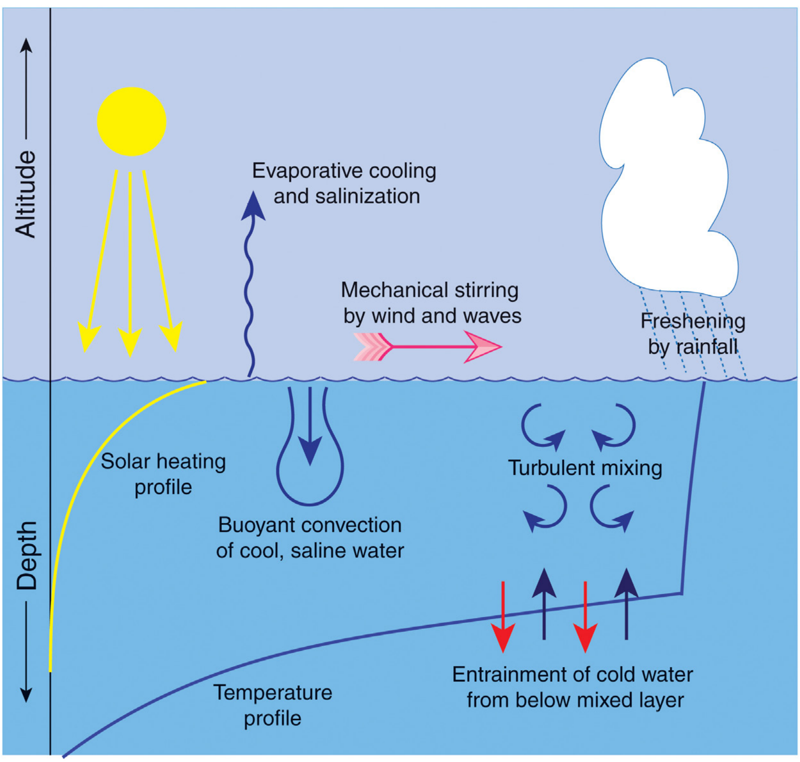

FIGURE 7.4 **Important mixed-layer processes.**

on the rate of buoyancy generation and the rate at which kinetic energy is supplied to the ocean surface by winds. If the surface is cooled very strongly, such as at high latitudes during fall and winter, then cold, dense water is formed near the surface at a rapid rate and buoyancy forces will drive convection, with sinking of cold water and rising of warmer water in the mixed layer. When the surface is cooled only weakly or actually heated, such as during summer when surface solar heating rates are greatest, then the generation of mixing by buoyancy is less and the mixed layer will become thinner and warmer. Buoyancy can be generated by the effect of evaporation on the surface salinity, even when surface temperatures are increasing with time. The density increase associated with increasingly saline surface waters can balance or overcome thermal stratification and encourage mixing. Rainfall represents an input of fresh water at the surface, which acts to decrease the density of the surface waters. Winds blowing over the ocean waves transfer kinetic energy to the water, which results in turbulent water motion as well as mean ocean currents. The supply of turbulent kinetic energy to the upper ocean by winds can induce mixing, even in the presence of stable density stratification. If the intensity of turbulence in the mixed layer is great enough, cool and dense water can be entrained into the mixed layer from below. This implies a downward heat transport, which cools and deepens the mixed layer.

The heat, momentum, and moisture exchanges between the atmosphere and the ocean are accomplished through contact of the atmospheric boundary layer with the mixed layer of the ocean. Storage and removal of heat from the ocean on time scales of less than a year are confined to the mixed layer over much of the ocean. The depth of the oceanic mixed layer varies from a few meters in regions where subsurface water upwells, as along the equator and in eastern boundary currents, to the depth of the ocean in high-latitude regions where cold, saline surface water can sink all the way to the ocean bottom. Regions where the mixed layer is deeper than 500 m constitute a small fraction of the global ocean area, however. In general, as one would expect, the mixed layer is thin where the ocean is being heated and thick where the ocean gives up its energy to the atmosphere. The global-average depth of the mixed layer is about 70 m. The mixed layer responds fairly quickly to changes in surface wind and temperature, whereas the ocean below the mixed layer does not. The thermal capacity of the mixed layer is the effective heat capacity of the ocean on time scales of years to a decade, and is about 30 times the heat capacity of the atmosphere (Chapter 4).

Figures 7.5a,b show the global distribution of mixed layer depth, temperature and salinity for January and July. During the summer the mixed layer is relatively shallow, especially in the Northern Hemisphere, but it becomes deeper in the winter, as expected, but particularly so in several locations. In the North Atlantic, the mixed layer becomes very deep in winter and these are the regions where deep water can be formed. During July (winter), in

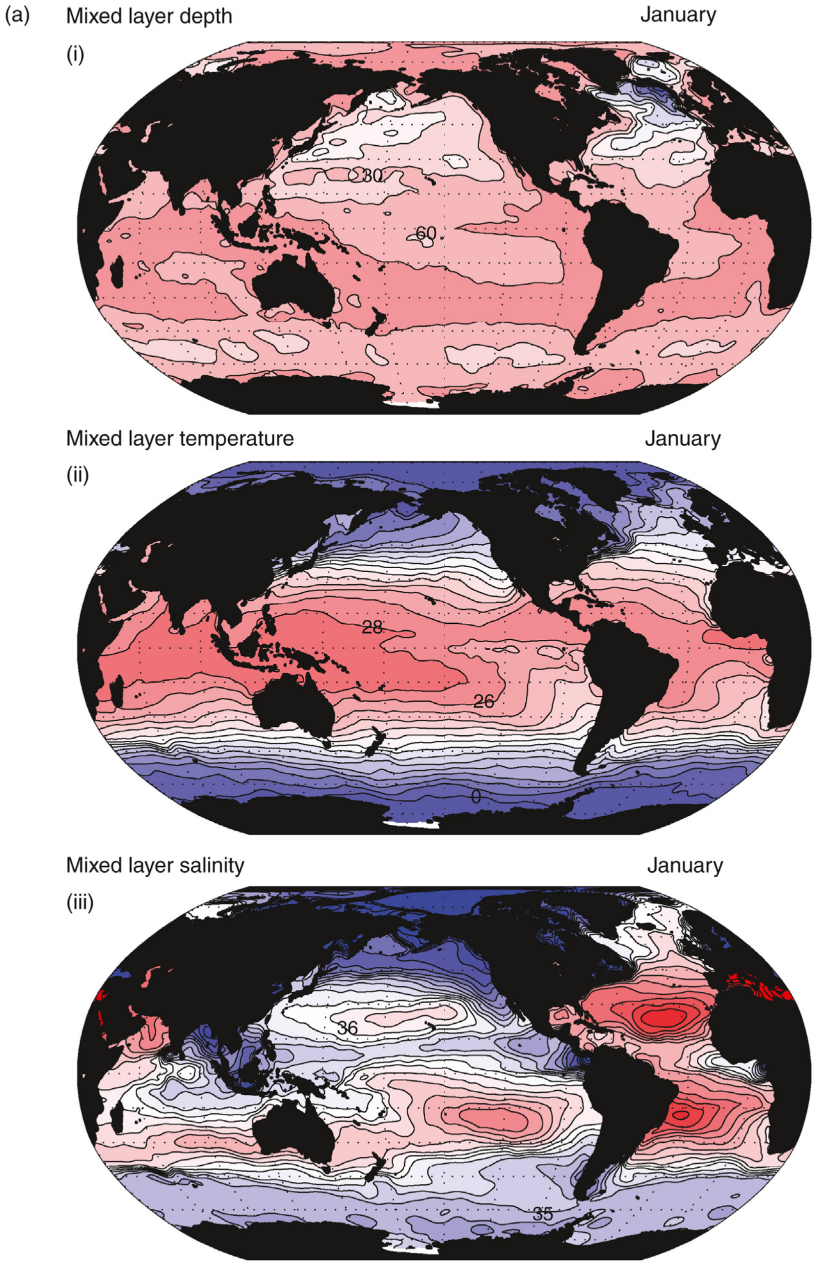

FIGURE 7.5 (a) (i) Map of the depth of the mixed layer for January. Contour interval is 30 dbars, with the reddest color more shallow than 30 dbar and the darkest blue colors the deeper mixed layer depths. (ii) Map of the potential temperature averaged over the mixed layer. Contour interval is 2°C, and warmest contour is 28°C. (iii) Map of salinity averaged over the mixed layer. Contour interval is 0.25‰ and saltiest contour in the Atlantic is 37.25‰. Data from MIMOC.

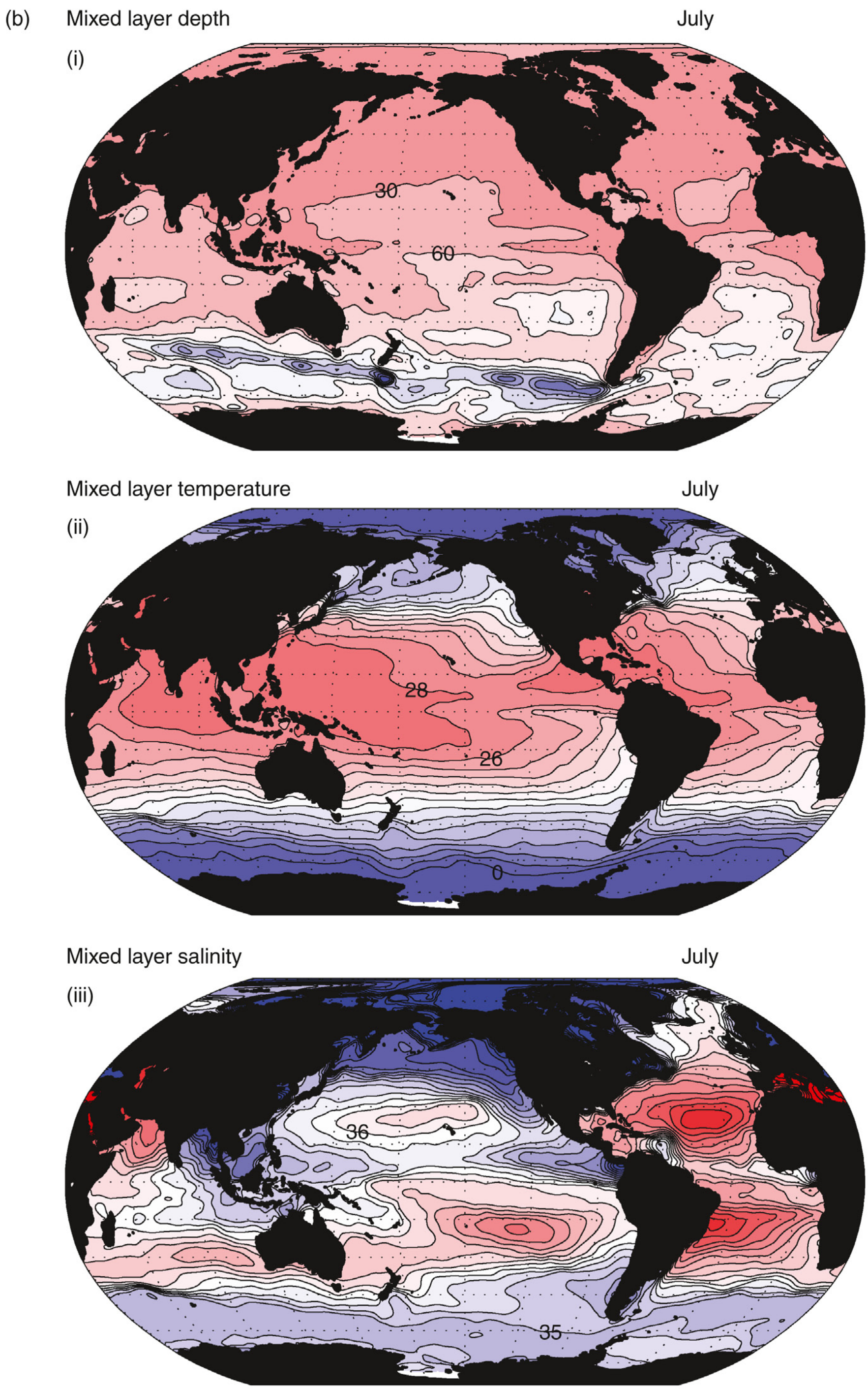

FIGURE 7.5 (*Continued*) (b) Same as Fig. 7.5a, except for July.

the Southern Hemisphere, a band of deepened mixed layer extends from the Indian Ocean to the eastern Pacific in the vicinity of the entrance to the Drake Passage off southernmost South America. This deepened mixed layer marks the subduction of cold, relatively fresh water into intermediate depths that can be seen in Fig. 7.2. The mixed layer does not become as shallow in summer in the Southern Hemisphere because the strong westerlies persist year-round there (Figs 6.4 and 6.19), and continue to stir the mixed layer in summer. As one would expect, the mixed-layer temperature and salinity are very similar to the surface values. The Atlantic is the saltiest ocean. Because of the geometry of the Atlantic Ocean and surrounding continents, some of the evaporation that occurs in the subtropical Atlantic is exported by the atmosphere to other ocean basins, mostly because of westward vapor transport across the Isthmus of Panama and southward transport toward the Southern Ocean along 30°S. As the climate warms this natural export of freshwater will strengthen with the increased moisture of the air, and the subtropical Atlantic will get even saltier.

Figure 7.6 shows examples of the seasonal cycle of the mixed layer from the North Atlantic and Southeast Pacific. The mixed layer is warmest and

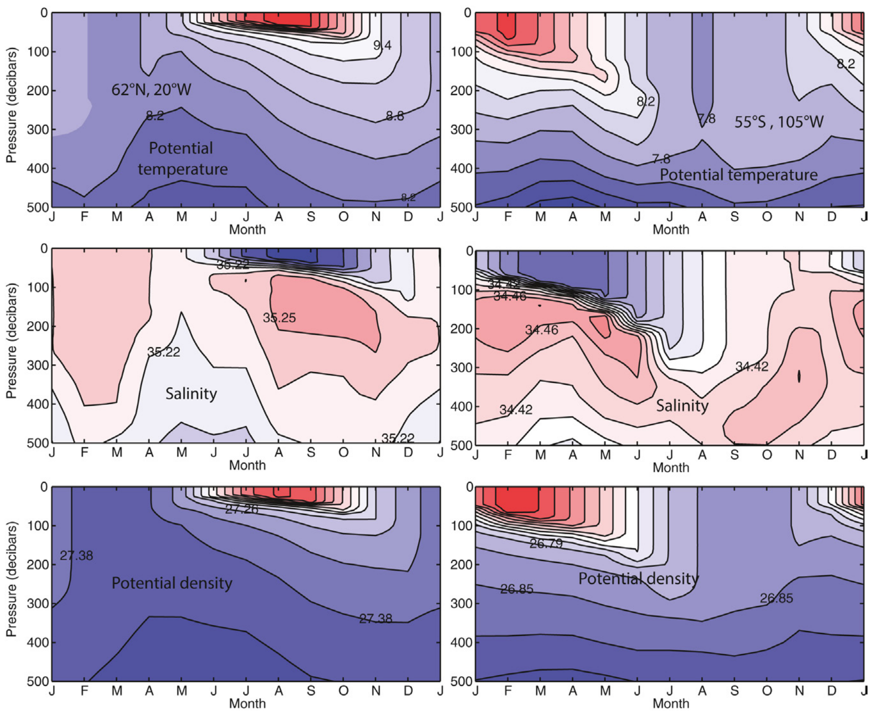

FIGURE 7.6 **Plots of potential temperature, salinity and potential density as functions of depth and month of the year in the far North Atlantic (62°N, 20°W) and in the South–West Pacific (55°S, 205°W).**

thinnest in late summer near the end of the period of greatest insolation and least intense stirring of the ocean by winds. In late summer, the surface begins to cool, the storminess increases, and the mixed layer deepens and cools. The mixed layer continues to deepen and cool throughout the winter, and by the end of winter, may extend to a depth of several hundred meters and merge smoothly into the permanent thermocline. During most of the rest of the year a seasonal thermocline with steep temperature gradients links the permanent thermocline with the base of the mixed layer. In spring and summer this seasonal thermocline develops and the mixed layer becomes thinner and warmer. Seasonal variations in temperature are confined primarily to the mixed layer and the seasonal thermocline, so temperatures at depths below the deepest extent of the mixed layer experience little seasonal variation. At both these locations, a fresh water layer caps the more saline water below during summer, but during winter salt is mixed upward, making a layer of cold, salty, and very dense water near the surface, especially in the far North Atlantic. These conditions allow dense water to be formed that can sink below the surface and fill the deeper layers of the ocean. Anomalies that are generated by weather or climate events in the winter can be sealed under the strong density stratification of the thin summer mixed layer, and then emerge again to influence the atmosphere in winter.

7.4 THE WIND-DRIVEN CIRCULATION

The transfer of momentum from winds to ocean currents plays a critical role in driving the circulation of the ocean. This is particularly true for the currents near the ocean surface. The wind stress acting on the ocean looks very much like the near-surface winds (Fig. 7.7). In the subtropics, especially in the winter hemisphere, the trade winds apply an equatorward and westward stress to the ocean. In the mid-latitudes, these stresses are reversed, and apply an eastward and poleward force on the surface. In the equatorial East Pacific and the Atlantic during July, the wind stresses are westward and these act to drive upwelling and colder SSTs there (Section 7.5).

The global distribution of large-scale surface ocean currents is shown in Fig. 7.8. These currents are not directly observed, but are inferred from observations of ocean surface height, wind stress, and SST. The basic data have a spatial resolution of 1° in latitude and longitude and many fine scale features are not represented. In the trade wind regions, the near-surface currents flow approximately 90° to the right of the applied wind stress vector in the Northern Hemisphere, and the reverse in the Southern Hemisphere. This is in general accord with the Ekman Theory, and will be introduced in Section 7.5. Near the equator in the Pacific, the current is

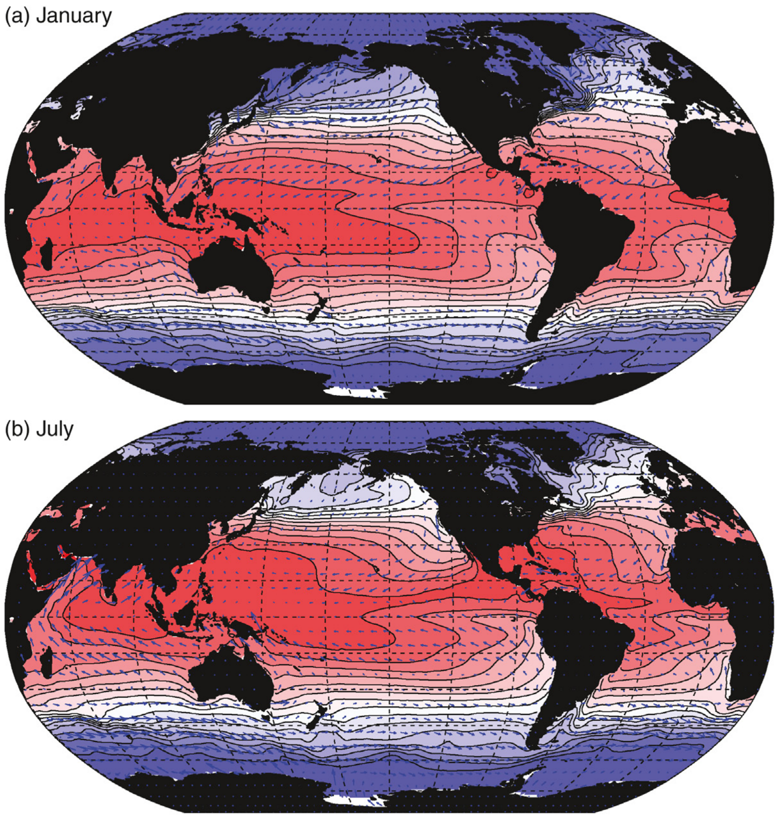

FIGURE 7.7 **Wind stress vectors and sea surface temperature (SST) for (a) January and (b) July.** *Data from ERAI.*

westward, but north of the equator the North Equatorial Countercurrent flows toward the east, most strongly in northern summer. In the Southern Ocean, the current flows generally eastward in a circumpolar current that squeezes through the Drake Passage between South America and the Antarctic Peninsula. The surface currents are arranged in coherent patterns, with large circulations called *gyres*, occupying the major ocean basins with currents circulating in the same direction as the applied wind stress.

The surface currents in the Indian Ocean show a strong response to the monsoonal wind reversal (Fig. 7.9). During January, the currents are westward north of the equator, which then reverse to eastward in July. Along the coast of Africa and Arabia, the southerly summer winds drive a northward current, which also drifts offshore to cause upwelling of cooler water along the coast. As a consequence of the upwelling and the stronger winds in summer, the SST is warmer in January than in July.

7.4.1 Western Boundary Currents

Some of the most visible current structures are the large clockwise circulations in the northern Pacific and Atlantic oceans. Along the western boundaries of the Pacific and Atlantic Ocean basins strong poleward-flowing currents exist in a narrow zone very near the continents. These currents are called the Kuroshio and the Gulf Stream, respectively, and may be referred to generically as western boundary currents. Western

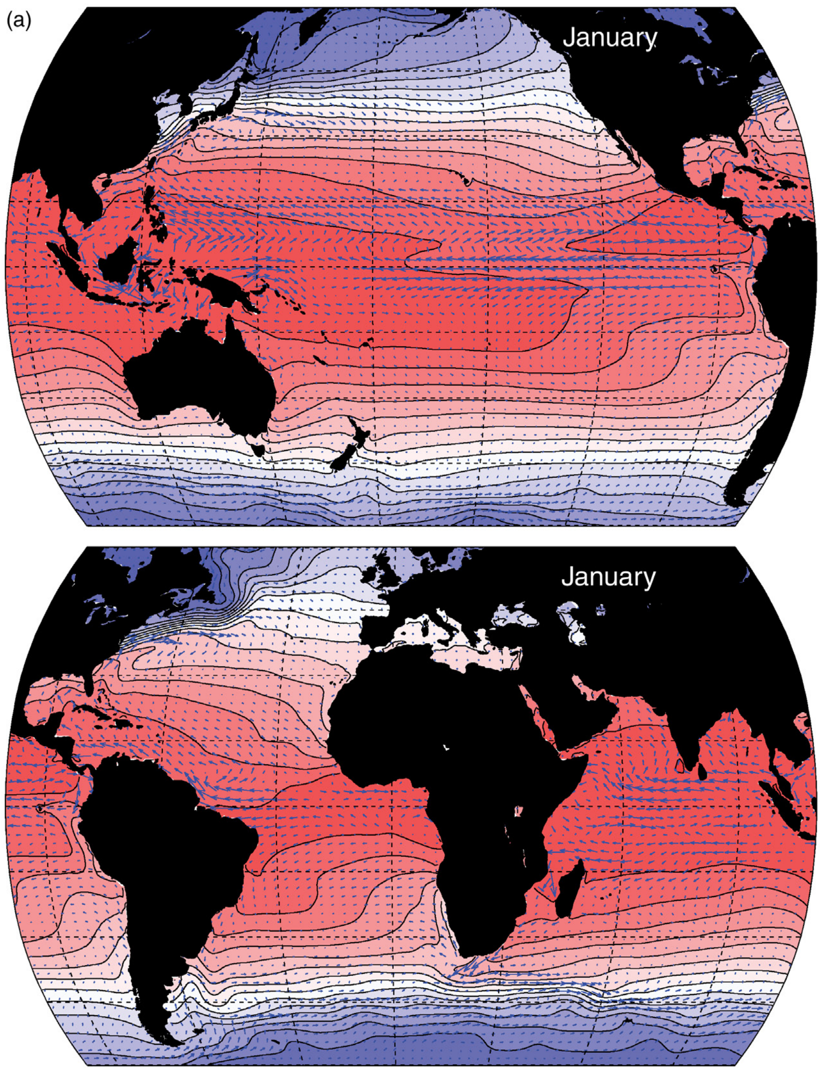

FIGURE 7.8 (a) Surface ocean current vectors and SST for January.

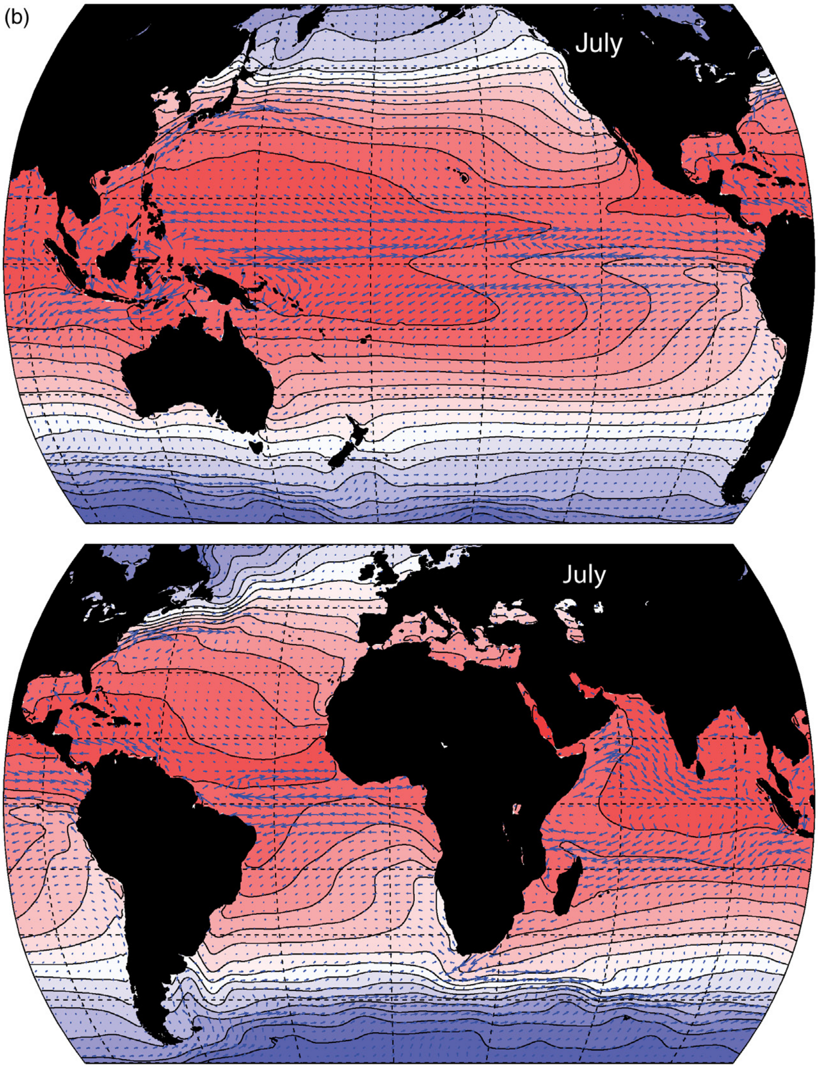

FIGURE 7.8 (*Continued*) (b) Surface ocean current vectors and SST for July. *Current vectors are from the OSCAR data set.*

boundary currents also occur in the Southern Hemisphere along South America (Brazil Current) and along Africa (Agulhas Current). They are generally less sharply defined and extensive in the Southern Hemisphere, perhaps because of the different ocean geometry that allows the Antarctic circumpolar current to flow unimpeded in a continuous eastward current at about 60°S. Western boundary currents carry warm water from the tropics to middle latitudes. The speed of these currents may exceed 1 m s^{-1},

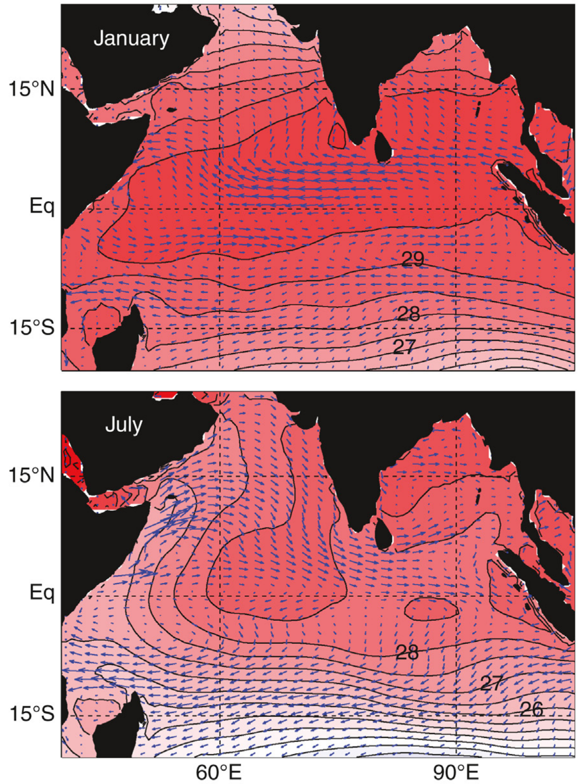

FIGURE 7.9 **Surface ocean current vectors and SST (°C) in the Indian Ocean region showing the seasonal reversal of currents associated with the monsoon, and the cooling of water during the summer monsoon associated with the upwelling along the coast of Africa and the greater surface winds during July.**

which is quite fast for an ocean current. With the possible exceptions of the Antarctic circumpolar current and some zonal equatorial currents, these currents are the closest oceanographic analogs to the jet streams of the atmosphere, although they flow mostly poleward rather than eastward. The return flow of water from mid-latitudes to the equator is much more gradual and occurs in a broad expanse across the center of each basin.

Figure 7.10 shows the surface current and SST in the western Atlantic region. Currents flow poleward along the western margin of the Atlantic, flowing along the South American coast into the Gulf of Mexico, through the Strait of Florida, and up the East Coast of North America. Around Cape Hatteras, they leave the coast and turn eastward, following the gradient of SST toward the far North Atlantic. Unlike the atmospheric jet streams that generally show a strong seasonal signature, the Gulf Stream continues with almost equal strength in the summer and winter.

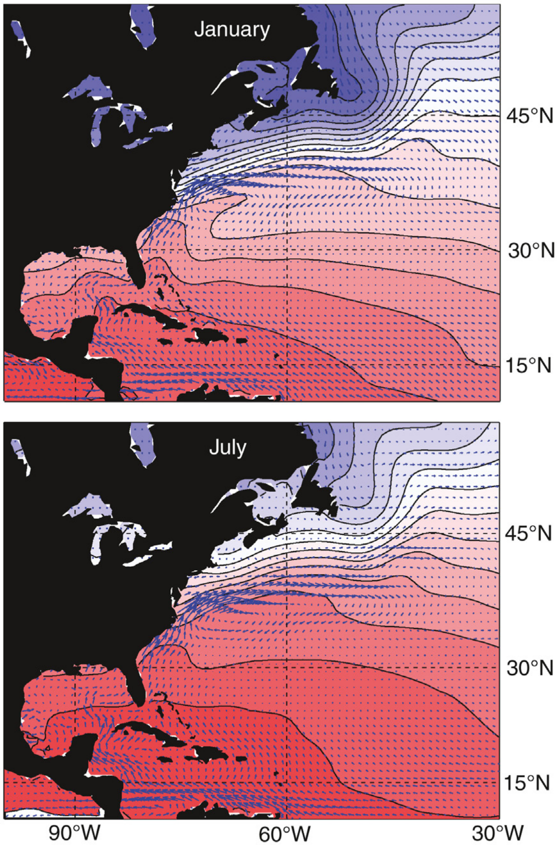

FIGURE 7.10 **Surface currents and SST in the western Atlantic Ocean region.**

The vertical structure of the Gulf Stream is illustrated in Fig. 7.11, which shows the geostrophic current and potential temperature in the cross-section along 65°W. The geostrophic current is obtained by assuming that the pressure surfaces are level at 2000 dbar, then computing the height above the 2000 dbar surface using the hydrostatic relationship (7.1).

$$z(p) - z(2000\,\mathrm{dbar}) = \int_{p}^{2000} \frac{1}{\rho g}\, dp \tag{7.1}$$

The density ρ depends on temperature, salinity, and pressure, and g is the acceleration of gravity. Once the heights of the pressure surfaces are known, the currents can be determined from the geostrophic approximation (Holton and Hakim, 2012).

$$u_g = -\frac{g}{f}\frac{\partial z}{\partial y}\bigg)_p \qquad v_g = \frac{g}{f}\frac{\partial z}{\partial x}\bigg)_p \tag{7.2}$$

The *Coriolis parameter* ($f = 2\Omega\sin\phi$) measures twice the local vertical component of the rotation rate (Ω) of Earth.

The current has a monthly mean speed of more than 0.5 m s^{-1} and is geostrophically balanced with the sloping thermocline, which implies temperature and density variations along lines of constant pressure. The eastward current at this longitude extends about 1000 m deep to the base of the thermocline and is slightly stronger and about 2° farther north during January than August. Even though averages over a month tend to blur out the sharpness of the jet structure, the current is still narrow; and the width over which the current exceeds half of its maximum strength is about 150 km.

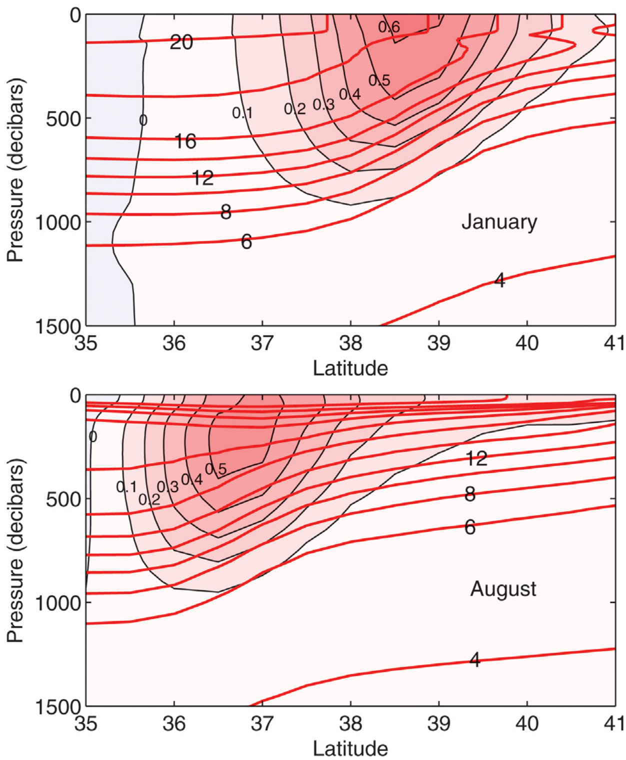

FIGURE 7.11 **Eastward geostrophic ocean current (black contours and shading) and potential temperature isolines (red contour lines) for January and August along a meridional cross-section at 65°W where the Gulf Stream travels approximately eastward.** Current data are in meters per second and temperatures in degree Celsius. Temperature, pressure and salinity data from MIMOC were used to compute the density.

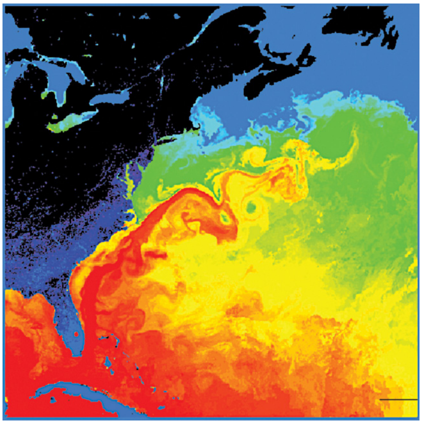

FIGURE 7.12 **SST in the Northwestern Atlantic showing meanders and rings in the Gulf Stream.** Red is warm and blue is cold. *Courtesy of Dr. Otis Brown.*

The Gulf Stream current is not straight or steady, but breaks down into meanders and rings, and eventually loses a clear identity as the flow expands eastward and northward across the basin (Figs 7.12 and 7.13). The current itself and these eddy structures are much smaller than the corresponding jets and eddies in the atmosphere by a factor of about 100.

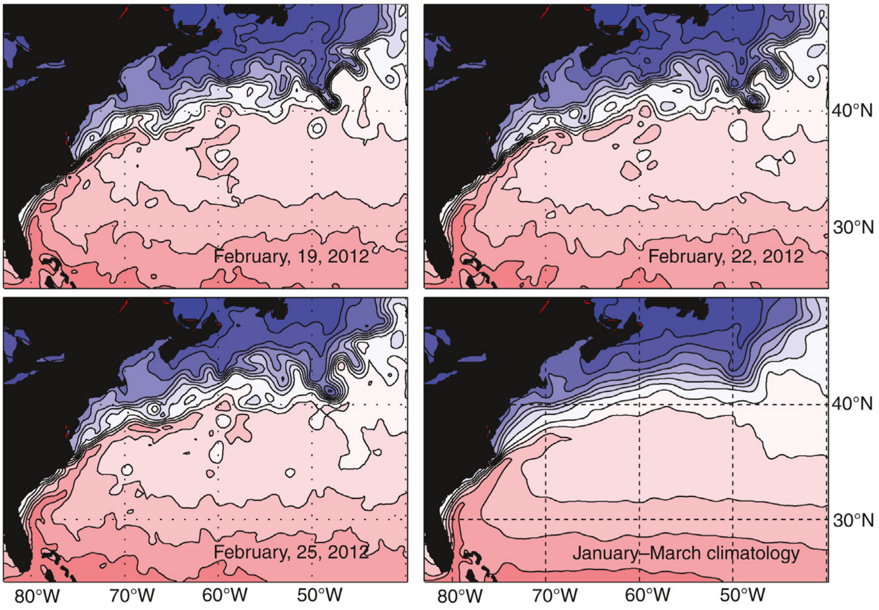

FIGURE 7.13 **Daily sea surface temperature in the Western Atlantic for 3 days in February 2012, plus the 3-month January–March climatology for SST.** *Data from NOAA OI SST high-resolution data set.*

The baroclinic *Rossby Radius of Deformation* is smaller in the ocean as is also the scale of maximum baroclinic instability, whereby eddies in the Gulf Stream derive their energy, as in the atmosphere. Some of the eddy structures in Fig. 7.13 are related to topographic features in the ocean bottom. On each of the three days shown, a vortex with an expression in cold SST exists near 43°N–50°W, and this feature also shows up in the climatology of January through March. This feature is associated with the subsurface topographic feature of the Grand Banks, where the cold Labrador Current from the north meets the Gulf Stream coming from the west and south. Downstream of the Grand Banks at about 43°N–47°W one can see the incorporation of a warm ring of water into colder surroundings over the course of February 19–25, 2012. Further west at about 38°N–68°W, one can see cold water wrapping around a warm intrusion, leading to a warm ring embedded within colder water on February 25, at about 39°N–67°W.

The poleward flux of warm water in the Gulf Stream and Kuroshio currents has a profound effect on the SST and climate of the land areas bordering the oceans, especially the lands immediately downwind of the oceans. The annual average SST shows a strong gradient in the North Atlantic Ocean aligned approximately with the mean position of the Gulf Stream (Fig. 7.13). This strong temperature gradient extends northeastward from the Mid-Atlantic coast of North America to the vicinity of Spitsbergen north of the Scandinavian Peninsula. Some of the heat carried northward by the Gulf Stream is picked up by the Norwegian Current and carried into the polar latitudes. As a result, at middle and high latitudes, the eastern Atlantic is much warmer at the surface than the western Atlantic Ocean and the Pacific Ocean. This asymmetry in the Atlantic sea surface temperature contributes to the milder winter climates of western European land areas compared to eastern North American land areas at the same latitude. Another major contribution to this climate asymmetry is the southward and eastward advection of cold air temperatures above North America in winter. These stationary waves in the atmosphere are caused by mountain ranges and the east–west SST variations.

7.4.2 Eastern Boundary Currents

Also important for climate are eastern boundary currents, which occur in tropical and subtropical latitudes at the eastern margins of the oceans. These currents are generally cold, shallow, and broad in comparison to western boundary currents. The names given to the eastern boundary currents in these geographic areas are the California Current off North America, the Peru or Humboldt Current off South America, the Canary Current off northern Africa, and the Benguela Current off southern Africa. In each

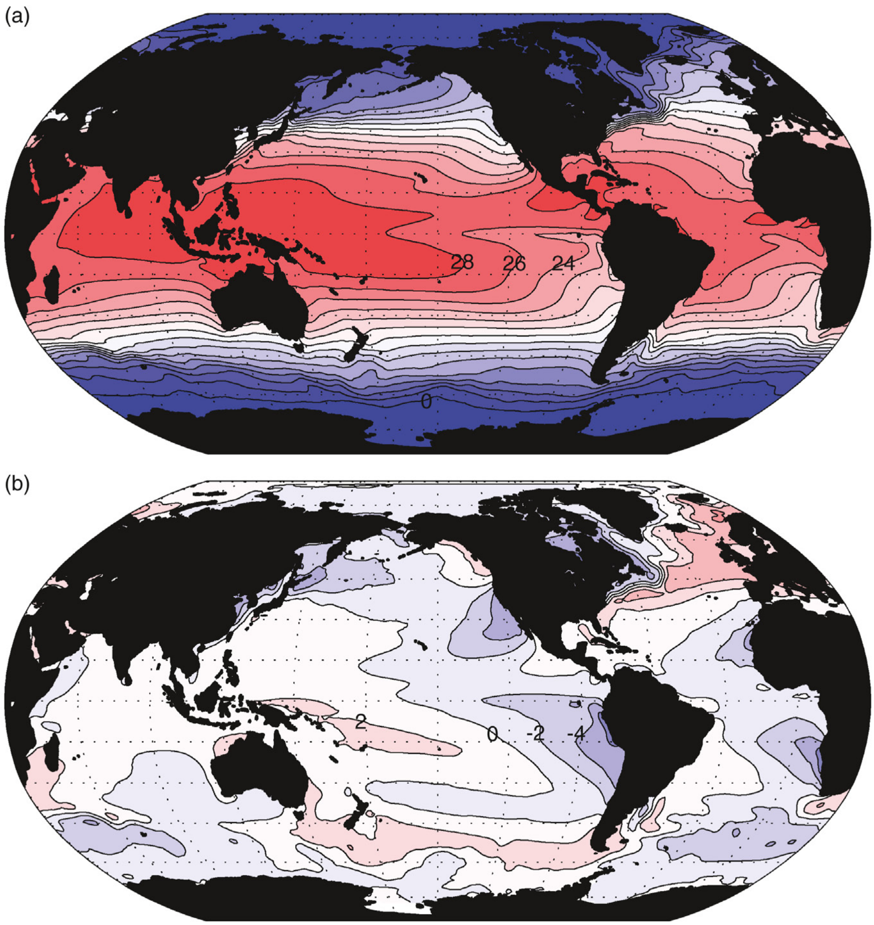

FIGURE 7.14 **(a) Annual mean SST and (b) deviation of the annual mean SST from its zonal average at each latitude Units are ˚C.** *Data from NOAA OI SST data.*

of these regions, a wind-driven current flows along the coast toward the equator and then turns westward toward the center of the basin. These currents are associated with cold SST, which can be illustrated by plotting the deviation of SST from its zonal average (Fig. 7.14). The SSTs in the subtropics to the west of the continents in the Atlantic and Pacific Oceans are much colder than the zonal average at the same latitude. The coldest water occurs very near the coast and extends westward and equatorward into the oceans.

Upwelling of cold subsurface water, which is driven by alongshore or offshore winds in these regions, produces the low SST near the coast. These low-level winds are associated with the surface high-pressure systems in the atmosphere above the eastern subtropical ocean areas, as shown in Fig. 6.19. The wind systems and the associated currents and cool SSTs are best developed during the summer in the Northern Hemisphere, and are more nearly year-round phenomena in the Southern Hemisphere. It is believed that the geometries of the coastlines in the two oceans, and in particular the northwest-southeast slopes of the coastlines of South America and Africa, cause the eastern boundary currents in the Southern Hemisphere to be better developed and to extend to the equator and then westward along the equator. The cooler than average SST in eastern boundary current regions is often associated with atmospheric subsidence and persistent low stratocumulus cloud and fog (Fig. 3.21).

7.5 THEORIES FOR WIND-DRIVEN CIRCULATIONS

7.5.1 The Ekman Layer, Wind-Driven Transport, and Upwelling

Due to the rotation of Earth, the frictional component of the vertically integrated transport of water in the surface layer of the ocean is not in the direction of the applied wind stress, but 90° to the right of it in the Northern Hemisphere and 90° to the left of it in the Southern Hemisphere. This wind-driven near-surface water transport plays a critical role in determining relatively cold surface temperatures in the eastern boundary current regions and along the equator, and it also plays an important role in driving the subtropical gyres that feed the western boundary currents.

To show the relationship between wind stress driving, currents, and transport, we may consider a homogenous ocean of constant density and pressure, and assume that it is driven by a uniform wind stress with eastward component, τ_x and northward component, τ_y. We seek a steady solution, in which frictional stresses and Coriolis accelerations are in balance (Vallis, 2006).

$$fv = -\nu \frac{d^2 u}{dz^2} \tag{7.3}$$

$$fu = \nu \frac{d^2 v}{dz^2} \tag{7.4}$$

The frictional forces have been described such that the frictional stress is proportional to the shear of the current velocity times a momentum diffusion coefficient, ν. Thus, the specified wind stress enters as a boundary condition on the current shear at the surface, and we assume that the current goes to zero at large depths, so that the boundary conditions on (7.3) and (7.4) are

$$\left.\begin{aligned} \nu\frac{du}{dz} &= \frac{\tau_x}{\rho_0} \\ \nu\frac{dv}{dz} &= \frac{\tau_y}{\rho_0} \end{aligned}\right\} \quad \text{at } z=0;\ u=v=0 \text{ at } z \to -\infty \tag{7.5}$$

where ρ_o is the density of the seawater and assumed constant. The solution for the velocities under these conditions is:

$$u_E = \frac{e^{z/z_E}}{\rho_0\sqrt{f\nu}}\left\{\tau_y \cos\left(\frac{z}{z_E}+\frac{\pi}{4}\right)+\tau_x \cos\left(\frac{z}{z_E}-\frac{\pi}{4}\right)\right\} \tag{7.6}$$

$$v_E = \frac{e^{z/z_E}}{\rho_0\sqrt{f\nu}}\left\{\tau_y \cos\left(\frac{z}{z_E}-\frac{\pi}{4}\right)-\tau_x \cos\left(\frac{z}{z_E}+\frac{\pi}{4}\right)\right\} \tag{7.7}$$

where, $z_E = \sqrt{\dfrac{2\nu}{f}}$ is the Ekman Depth and z is measured positive upward.

The steady solution (7.6)–(7.7) describes the Ekman spiral (Fig. 7.15). The current vector has its maximum magnitude at the surface where it is directed at an angle of $\pi/4$ (45°) to the right of the wind stress vector in the Northern Hemisphere ($f > 0$). The current vector turns toward the right with increasing depth, and its magnitude decreases exponentially with depth. The magnitude of the current decreases by a factor of e^{-1} for every increase of depth equal to $z_E = \sqrt{2\nu/f}$. An appropriate value of $\nu = 30\ \mathrm{m^2\,s^{-1}}$ gives an Ekman depth of ~ 800 m. If we integrate the currents over the depth range in which the currents are significant, we obtain the integrated transport in the Ekman layer.

$$U_E = \int_{-\infty}^{0} u_E\, dz = \frac{\tau_y}{\rho_0 f}; \quad V_E = \int_{-\infty}^{0} v_E\, dz = \frac{-\tau_x}{\rho_0 f} \tag{7.8}$$

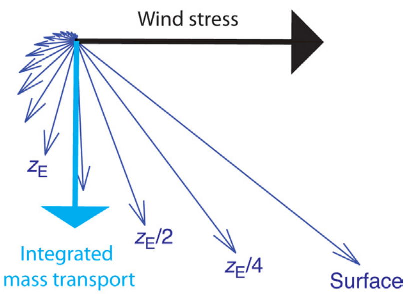

FIGURE 7.15 **Schematic of the Ekman Layer solution (7.6, 7.7 and 7.8) showing the applied wind stress, the current vectors at various depths, and the mass transport integrated through depth.** Current vectors are shown at depth intervals of $z_E/4$.

The net horizontal water transport in the Ekman layer is directed at a 90° angle to the right of the applied wind stress in the Northern Hemisphere, so that if the wind stress is toward the east at the surface ($\tau_x > 0$), the Ekman layer transport is toward the south ($V_E < 0$). If a westward wind stress is applied near the equator, the Ekman layer transport will be northward in the Northern Hemisphere and southward in the Southern Hemisphere, because of the change of sign of f at the equator, so that a net divergence of surface flow will be generated. Conservation of mass requires upwelling along the equator to balance the Ekman layer transport away from the equator. The cold tongue of SST in the eastern equatorial Pacific Ocean during July (Fig. 7.9) is caused largely by this wind-driven upwelling. The cold SST anomalies associated with the eastern boundary currents (Fig. 7.14) are associated with offshore Ekman transport driven by equatorward alongshore surface winds. Offshore Ekman transport near an ocean boundary requires upwelling to replace the exported water. Since water temperatures decrease with depth, upwelling is generally accompanied by cold SST.

Wind stress driving can also cause vertical motions in the open ocean away from boundaries and the equator, if the wind stress has spatial gradients. If we consider the mass continuity equation for an incompressible fluid

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0 \tag{7.9}$$

and integrate it over the depth of the Ekman layer, we can derive a relationship between the applied wind stress and the vertical motion at the base of the Ekman layer.

$$w_{\mathrm{E}}(-\infty)-w_{\mathrm{E}}(0)=\int_{-\infty}^{0}\left(\frac{\partial u_{\mathrm{E}}}{\partial x}+\frac{\partial v_{E}}{\partial y}\right)dz \tag{7.10}$$

Utilizing (7.8) in (7.10) and assuming that the vertical current vanishes at the surface yields an expression for the vertical velocity at the bottom of the Ekman layer in terms of the wind stress applied at the surface. We obtain,

$$w_{\mathrm{E}}(-\infty)=\frac{\partial}{\partial x}\left(\frac{\tau_y}{\rho_0 f}\right)-\frac{\partial}{\partial y}\left(\frac{\tau_x}{\rho_0 f}\right)=\vec{k}\bullet\nabla\times\left(\frac{\vec{\tau}}{\rho_0 f}\right) \tag{7.11}$$

where $\vec{\tau}=\vec{i}\,\tau_x+\vec{j}\,\tau_y$ and $\vec{i}$, $\vec{j}$, and $\vec{k}$ are unit vectors in the eastward, northward, and upward directions respectively. The vertical velocity at the base of the Ekman layer in the open ocean is thus seen to be proportional to the curl of the wind stress vector divided by the Coriolis parameter. Where lateral boundaries are present, the dependence of upwelling on the wind stress is more complex, but wind stress near boundaries can produce large upwelling, even without significant wind stress curl.

7.5.2 Sverdrup Flow and Western Boundary Currents

To understand the large-scale response of the ocean to wind stress forcing it is useful to consider the balance of vorticity in the ocean. *Vorticity* is the curl of the velocity vector and is a measure of the local rotation of the fluid. For the large-scale motions of the atmosphere and the ocean it is the vertical component of absolute vorticity that is of most interest.

$$\zeta_a = 2\Omega \sin\phi + \vec{k} \bullet \vec{\nabla} \times \vec{V} = f + \zeta_r \tag{7.12}$$

The absolute vorticity is the sum of planetary vorticity (f), which is associated with the rotation of Earth, and relative vorticity (ζ_r), which is associated with the fluid motion relative to the surface of Earth. For flow without friction, the absolute vorticity remains constant, unless a parcel of fluid changes its shape. If a parcel of fluid maintains its shape while moving equatorward to a latitude where Earth's rotation is less, then the fluid parcel must exhibit a change in relative vorticity in order to maintain a constant absolute vorticity. Stretching of fluid parcels along the direction of the rotation vector will cause the absolute rotation rate to increase.

The famous oceanographer Harald Ulrik Sverdrup showed that in the interior of the ocean, an approximate balance exists between the meridional advection of planetary vorticity and the stretching of planetary vorticity by divergent motions,

$$\beta v = f \frac{\partial w}{\partial z} \tag{7.13}$$

where $\beta = \partial f / \partial y = 2\Omega \cos\phi / a$. If we integrate (7.13) from the bottom of the ocean to the bottom of the Ekman layer and use (7.11), we obtain

$$\beta V_I = f \vec{k} \bullet \nabla \times \left(\frac{\vec{\tau}}{\rho_0 f} \right) \tag{7.14}$$

where

$$V_I = -\int_{-D_0}^{-z_E} v \, dz \tag{7.15}$$

and it has been assumed that the Ekman layer is thin compared to the depth of the ocean, and that the vertical velocity is zero at the bottom of the ocean, where $z = -D_o$.

If we add the interior meridional transport velocity (7.14) to the Ekman layer meridional transport in (7.8) we obtain,

$$V_{\mathrm{I}} + V_{\mathrm{E}} = \frac{1}{\beta}\vec{k} \bullet \nabla \times \left(\frac{\vec{\tau}}{\rho_0} \right) \tag{7.16}$$

so that the total meridional mass transport is proportional to the curl of the wind stress.

If we consider the ocean circulation between the tropics and mid-latitudes, the wind stress varies from westward in the tropical easterlies to eastward in the mid-latitude westerlies (Fig. 7.7). Thus a wind stress curl opposite to the rotation of the Earth is applied to the ocean, and according to (7.16), we should expect the water transport in the ocean to be equatorward. Physically, the wind stresses are causing the water to rotate about a vertical axis in a direction that is opposite to the rotation of Earth. To maintain a steady state in the face of this application of anticyclonic rotation, water must drift toward lower latitudes. The reduction in absolute vorticity applied by the wind stress is thus expressed as a decrease in the planetary vorticity of fluid parcels as they drift equatorward, and a steady state with constant relative vorticity can be maintained.

According to (7.16), the meridional transport in the ocean will be equatorward everywhere, so long as the wind stress curl is opposite to Earth's rotation. How, then, can the conservation of mass and vorticity be jointly satisfied if the wind stress curl is everywhere negative? How does the water-transported equatorward return to high latitudes and close the circulation of mass and vorticity? The western boundary currents observed in the mid-latitude oceans are the solution to this dilemma.

A simple model can be constructed by adding a lateral diffusion term to the vorticity equation (7.13) that produces a steady gyre circulation with the northward flowing return current intensified along the western margin of the ocean basin, much like the observed northward flow is intensified in the western boundary currents. As the water flows poleward along this western margin, the planetary component of vorticity (f) increases because of its dependence on latitude. If the absolute vorticity of the fluid parcels were to be conserved, then their relative vorticity would have to change to compensate for the increase in planetary vorticity. By flowing along the western margin of the ocean basin in a narrow current, the poleward return flow of the wind-driven circulation is able to collect enough vorticity in the same sense as Earth's rotation to arrive at middle latitudes with a vertical component of absolute vorticity, near that of the planetary vorticity at these latitudes, so that the magnitude of the relative vorticity remains reasonably small and steady. In a simple linear model with lateral diffusion of momentum, the mechanism for collecting the necessary vorticity is through the lateral friction stresses, which generates vorticity of the proper sign, only along the western boundary of the ocean basin. Thus the warm, rapidly flowing western boundary currents of the Atlantic and Pacific Oceans are seen to be a response to the wind stress driving the bounded oceans of the Northern Hemisphere.

7.6 THE DEEP THERMOHALINE CIRCULATION

The term *thermohaline circulation* is used to denote that part of the oceanic circulation that is driven by water density variations, which are in turn related to sources and sinks of heat (thermo) and salt (haline). It is traditional in oceanography to organize the discussion of the oceanic circulation into separate wind-driven and density-driven components, although the circulation of the ocean is not a simple addition of the effects of these two types of forcing. Wind driving influences the sources and sinks of heat and salt for the ocean by transporting surface water from the tropics to latitudes where cooling and evaporation can increase its density to very large values. The heat transport associated with the thermohaline circulation affects the SST gradients that help to drive atmospheric winds. Therefore, wind driving and density driving of the oceanic circulation are very closely coupled and cannot be easily separated. However, it is generally true from a diagnostic perspective that wind driving is the strongest influence on currents near the surface, and variations in density are important for the flow at depth.

Slow circulations below the thermocline are driven primarily by density gradients in the deep ocean. These circulations are difficult to measure directly, since the currents associated with them are very weak, but their nature can be inferred from the distributions of trace constituents of seawater. Away from the surface, temperature and salinity of water masses change very slowly, so that the water masses and their origins can be inferred from the particular combination of temperature and salinity that characterizes them. These layers are often deep and the characteristics within them quite uniform, so that oceanographers refer to them as mode water (McCartney, 1982).

Most gases are soluble in water, so that the concentrations of particular gases can also be used to characterize water sources. The saturation concentration of a gas in seawater is the amount that would exist in solution at equilibrium, if seawater at a particular temperature and salinity were exposed to the gas. Saturation concentrations of gases in seawater increase as the water gets colder. For example, the saturation concentrations of oxygen and carbon dioxide in seawater at 0°C are about 1.6 and 2.2 times their values at 24°C, respectively. The concentration of oxygen in surface water is always slightly greater than its saturation value, probably as a result of efficient mixing of bubbles of air into the surface water and production of oxygen in surface waters by photosynthesis. When surface water sinks into the deeper levels of the ocean its source of oxygen is cut off, and the oxygen is slowly consumed by bacteria as they consume organic matter at depth. Therefore, one may use the depletion of the oxygen concentration below its saturation value as a measure of the time since the water has been at the surface.

Figure 7.16 shows the oxygen saturation versus depth and latitude in the Atlantic and Pacific oceans. In the North Atlantic Ocean we observe that high saturation values extend to great depths, and that these high values extend toward the Southern Hemisphere at depths below ~1500 m. We infer then that significant downwelling of water occurs in the North Atlantic and that this water sinks most of the way to the bottom of the ocean and then spreads southward. In the tropics, oxygen-depleted water exists just a few hundred meters below the surface, indicating that water has been below the surface for a long time. A stability barrier associated with the strong density stratification in the tropics prevents mixing of surface waters with the older water below. The distribution of oxygen saturation in the North Pacific Ocean is very different from that in the North Atlantic. In the North Pacific, we see no evidence of downwelling, and in fact the oxygen at depth is severely depleted, with saturations about 10–15% at latitudes and depths where the oxygen saturation is about 85% in the North Atlantic.

From the oxygen saturation alone we can infer that water from the surface sinks relatively quickly to the deep ocean in the North Atlantic, but that this does not occur in the Pacific. We cannot infer the full circulation of the deep ocean from oxygen alone, nor can we directly infer the subsidence rate, since the rate of oxygen depletion depends on the biological activity at depth, which in turn depends on the rate at which nutrients are supplied to them by deposition from above. The inferences from temperature, salinity, oxygen, and many other tracers suggest a deep-water circulation in the Atlantic in which a large mass of deep water is formed in the northern margin of the ocean, which then flows southward to fill a large fraction of the deep Atlantic (so-called North Atlantic Deep Water, NADW). This water rises toward the surface again in the vicinity of 60°S. Cold, but lower salinity water is formed in mid-latitudes of the Southern Hemisphere and wedges itself between the warm surface water and the North Atlantic deep water below. This sub-Antarctic mode water (SAMW) and closely related Antarctic intermediate water (AAIW) are more evident in the salinity distribution in Fig. 7.2, which shows relatively fresh water formed in southern mid-latitudes is subducted on the northern edge of the circumpolar current during winter as further evidenced by the deepened mixed layer there (Fig. 7.5b). AAIW is freshened from sea ice melting and is more closely identified with the deepened mixed layer, west of South

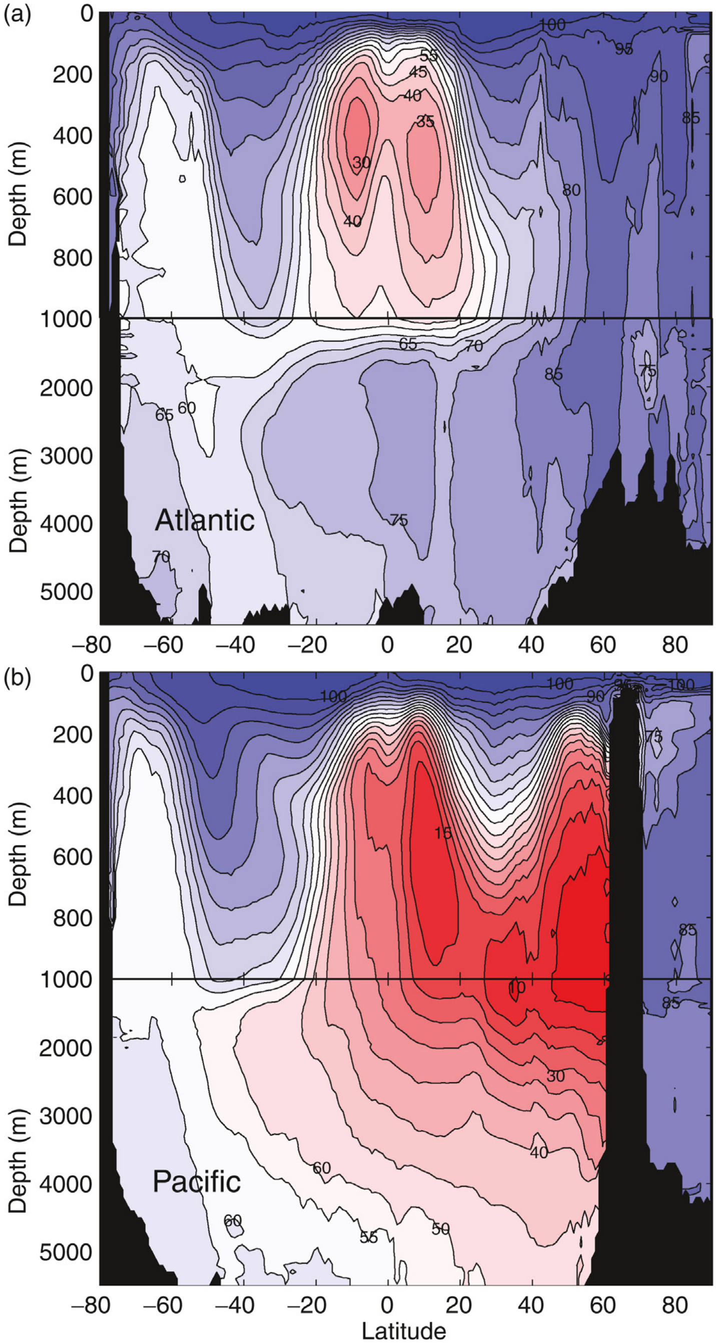

FIGURE 7.16 Oxygen saturation percentage for the (a) Atlantic and (b) Pacific oceans. Note that the vertical scale changes at 1000 m, since variations are much slower below this depth. *Data from NOAA.*

America. Bottom water is formed around Antarctica, especially in the Weddell Sea where seasonal ice formation adds salt to very cold seawater.

The mechanisms of deep-water formation in the North and South Atlantic are believed to be somewhat different. In the North Atlantic, warm, saline water flows poleward from mid-latitudes, where the Gulf Stream provides an important source of such water. This water is carried farther poleward into the Norwegian and Greenland Seas, where it is exposed to very cold atmospheric temperatures. The cooling of this saline water produces water that is dense enough to sink to great depths. The surface water in high latitudes of the Southern Hemisphere is relatively fresh, because of the excess of precipitation over evaporation in those latitudes, and no western boundary current exists to carry warm, saline water poleward, since the circumpolar current inhibits efficient transport of water from middle to polar latitudes in the southern oceans. Some saline water reaches the surface of the southern polar oceans by flowing southward at intermediate depths and rising at high latitudes in the south Atlantic. This water is also enhanced in nutrients, since it has spent some time at intermediate depths, where nutrients can be dissolved from falling detritus and photosynthetic organisms do not exist to consume the nutrients. In the waters around the Antarctic continent, which are the source regions for much of the Antarctic bottom water, the formation of very dense water is more dependent on sea ice production. Dense water formation is especially efficient in openings in the sea ice, or *polynyas*, where open water and very cold, dry air from come together and foster rapid sea ice production. This process is particularly efficient around Antarctica when cold offshore winds push the ice seaward and new sea ice is formed rapidly in the open water behind the northbound ice. When ice is formed from seawater, salt is rejected from the crystal structure, resulting in the formation of brine, which adds salt to the water immediately under the ice and thereby increases its density. This cold, saline water is dense enough to sink to the bottom of the Southern Ocean.

To infer the rate of downwelling it is necessary to use tracers with known decay times such as carbon-14 (^{14}C) or CFCs. Carbon-14, or radiocarbon is a radioactive isotope produced naturally in the atmosphere by cosmic rays and by the explosion of atomic bombs in the atmosphere. Since the rate of decay of radiocarbon is precisely known, the abundance of radiocarbon can be used to estimate how long water has been below the surface. The rate and spatial distribution of downwelling can also be inferred from transient trace gases such as chlorofluorocarbons (CFCs), which are man-made and have been introduced into the atmosphere starting in the middle of the twentieth century.

By combining the evidence available from tracers of seawater movement it has been convincingly shown that at the present time deep ocean water is formed only at high latitudes in the North Atlantic and Southern Ocean. Only in these locations can water of sufficient density be formed to sink to the deep ocean. From these two locations water spreads out at depth to fill the Pacific and Indian oceans, where the water gradually rises toward the surface. Since the North Pacific is the farthest from either of these two locations, the water at intermediate depths in the North Pacific is the "oldest" ocean water in the sense that it has been the longest time since this water was exposed to the atmosphere. The fact that the oldest water is not at the ocean bottom suggests that the deep water formed in the polar regions slowly rises elsewhere, as would be required by the conservation of water mass. The regions of the ocean where deep water can be formed constitute a small fraction of the total surface area of the ocean. For example, 75% of the ocean has potential density greater than 27.4, but only 4% of the surface water has a density that high. It is estimated that the time required to replace the water in the deep ocean through downwelling in the regions of deep-water formation is on the order of 1000 years. We may call this the turnover time of the ocean. The thermal, chemical, and biological properties of the deep ocean therefore constitute a potential source of long-term memory for the climate system on time scales up to a millennium. Some chemical properties of the ocean take longer than one turnover time to change significantly, so that the potential exists for ocean memory on time scales longer than the ocean turnover time.

Fig. 7.17 shows meridional cross-sections for the zonally averaged global ocean and for sections at 30°W in the Atlantic and 165°W in the Pacific. The turbulent mixing in the deep ocean is very weak and nearly adiabatic, so the flow of deepwater must be approximately along lines of constant potential density (isopycnals). Schematic arrows showing the hypothesized flow of deepwater are indicated for the global ocean in Fig. 7.17. Green arrows indicate the flow of North Atlantic Deep Water that is formed in the far North Atlantic Ocean and flows southward at intermediate depths. This water rises up again in the Southern Ocean, where the isopycnals slope upward and reach the mixed layer (Marshall and Speer, 2012). Antarctic Bottom Water is formed around the edge of Antarctica, especially in the Ross and Weddell seas, where the formation of ice during the winter season adds salt to the water below the ice, making very dense water. In the Southern Ocean the contours of oxygen saturation tend to follow the contours of potential density, supporting the schematic picture of the deep ocean circulation indicated by the colored arrows. The Southern Ocean is thus a key region for both forming bottom water and bringing deep water back to the surface.

The strongest ocean current is the Antarctic circumpolar current. The mass flow through the Drake Passage is about 125 Sverdrups ($125 \times 106\ m^3\ s^{-1}$) that flows eastward around Antarctica and passes through the Drake Passage between South America and Antarctica. It is driven by the very strong wind stresses associated with the mid-latitude jet in the atmosphere. The wind stress curl in the Southern Ocean drives upwelling poleward of the jet and downwelling equatorward of it. This

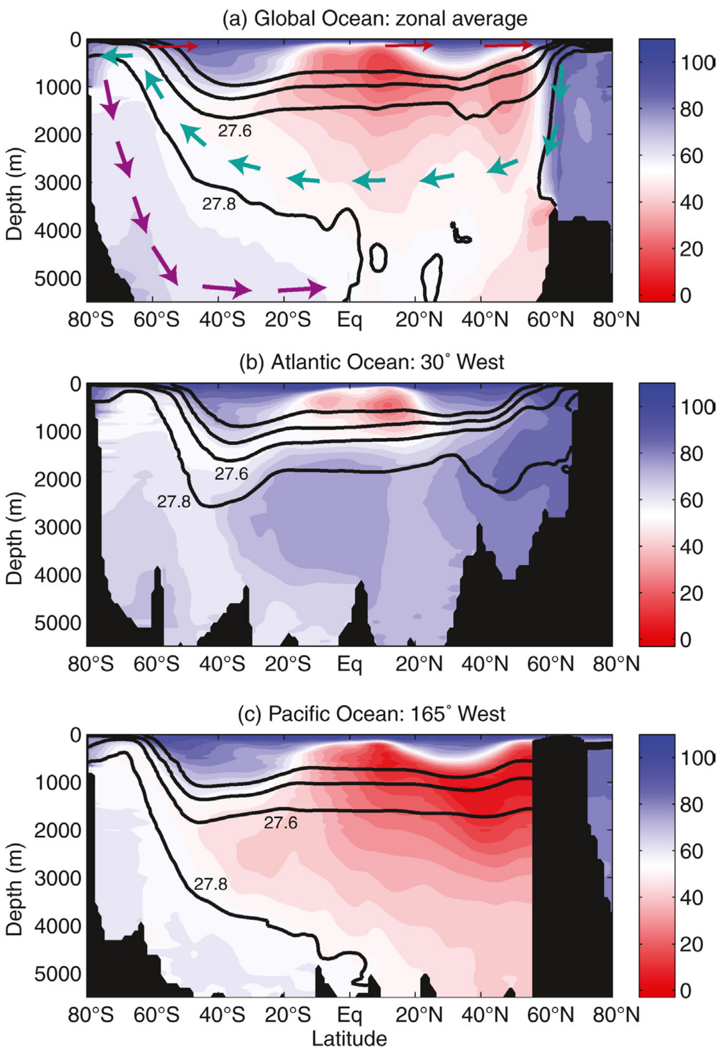

FIGURE 7.17 Meridional cross-sections for (a) the global ocean – zonally averaged, (b) along 30°W in the Atlantic Ocean, and (c) along 165°W in the Pacific Ocean. Colors indicate oxygen saturation in percent, solid contours are potential density minus 1000 as in Fig. 7.6. Contours are shown for 27.2, 27.4, 27.6 and 27.8 kg m^{-3}. Arrows indicate the flow of deep water, which tends to parallel contours of potential density. Purple vectors indicate the movement of Antarctic bottom water and green vectors indicate the movement of North Atlantic intermediate water. *Data are from NOAA.*

is consistent with the displacement of the contours of oxygen saturation, with old oxygen-poor water drawn upward poleward of the jet and a deep pool of oxygenated water equatorward of the jet (Fig. 7.17). This wind-driven vertical circulation also increases the temperature gradient and the slopes of the isopycnals, since dense water is drawn up on the poleward side and less dense surface water is pushed downward on the equatorward side. This strong density gradient fosters baroclinic instability of eddies in the ocean, and the Southern Ocean has very vigorous ocean eddies, much as in the mid-latitudes of the atmosphere, but with a horizontal scale of only about 100 km. These eddies produce fluxes of heat and potential vorticity that are important for ocean dynamics and tend to smooth out the meridional gradients that the wind-driven circulation generates.

The Southern Ocean is a key region for climate change, along with the North Atlantic and the tropical Pacific. The Southern Ocean seems to be the key region where deep water returns to the surface, whereas upwelling of less dense water is important in the tropical Pacific and deep water is formed in both the North Atlantic and the ocean around Antarctica. The circulation of the Southern Ocean is important for many aspects of climate variability and change. Since it is the primary region for return of deep water to the surface, the North Atlantic overturning circulation and the Southern Ocean upwelling are closely linked. The return of deep water to the surface means that warming of the Southern ocean will lag behind warming of the globe, and evidence of this effect will be seen in temperature trends calculated in Chapter 9. The circulation of the Southern Ocean interacts with the wind stresses. During an ice age, for example, the strongest wind stresses may move equatorward. If they move equatorward enough, then the importance of the Drake Passage for the circulation of the Ocean may be changed. The Southern Ocean may also be a key player in the carbon cycle variations that occur during ice ages. If sea ice extends equatorward along with the strongest winds, then ventilation of the deep ocean and the return of carbon to the surface may be suppressed, contributing to the reduction of atmospheric CO_2 that supports the ice ages. Reversing these changes may also contribute to the return of CO_2 to the atmosphere during interglacials, the warmer periods between the ice ages that define the large glacial–interglacial cycles of the past 800,000 years.

7.7 TRANSPORT OF ENERGY IN THE OCEAN

The general circulation of the ocean produces horizontal transport of energy from the tropics to the polar regions that is important for climate. However, it is not easy to measure this heat transport directly. It is difficult and expensive to obtain simultaneous current and temperature measurements from the surface to the bottom of the ocean. Such measurements require a ship or a large buoy and a cable with thermistors and current meters that extends from the surface to the ocean bottom, which is a distance of about 4 km on average. The spatial scales of the motions that are important for heat transport in the ocean are often small compared to the great expanse of an ocean basin, so that it is beyond our means to simultaneously measure current and temperature at enough spatial points and frequently enough in time to continuously monitor the product of velocity and temperature that produces most of the heat transport in the ocean. Attempts have been made to measure a series of profiles across a basin at a particular latitude, but these estimates must be assigned a rather large uncertainty.

An alternative to direct measurement of currents and temperatures is to infer the heat transport of the ocean from the energy balance of Earth or of the ocean. In Fig. 2.15, we inferred the annual mean northward heat flux by the ocean as a residual of the top-of-atmosphere and atmospheric heat budgets. This estimate indicates that the maximum meridional energy transport by the oceans in the Northern hemisphere is much smaller than the atmospheric energy transport, and occurs at a lower latitude, peaking in the tropical rather than the middle latitudes. A clear asymmetry can be seen between the Northern and Southern Hemispheres, which is likely related to the shape of the oceans and land areas in the two hemispheres. The poleward heat transport by the ocean in the Northern Hemisphere is about twice that in the Southern Hemisphere.

The oceanic energy flux can also be estimated from the energy balance at the surface of the ocean. Following (4.1) the energy balance at the surface can be written as (7.17).

$$\nabla \bullet \vec{F}_{\mathrm{o}} = R_{\mathrm{s}} - LE - SH - \frac{\partial E_{\mathrm{s}}}{\partial t} \tag{7.17}$$

The divergence of energy transport in the ocean can be estimated from the ocean surface energy balance if the surface net radiative heating, the evaporative cooling, the sensible cooling, and the energy storage in the ocean can be estimated. All of these terms are discussed in Chapter 4, and maps of the inferred oceanic flux divergence appear in Fig. 4.19. Estimates of meridional energy transport in the ocean obtained from (7.17) are in general agreement with estimates shown in Fig. 2.15, in that they show a maximum transport by the oceans at about 20ºN. The surface energy balance method can also provide estimates for individual regions and oceans (Trenberth et al., 2001). The estimates indicate that the Atlantic Ocean transports energy northward across the equator, whereas the Indian Ocean transports energy southward.

7.8 MECHANISMS OF TRANSPORT IN THE OCEAN

The meridional transport of energy in the oceans is clearly important for climate, but because the transports are not measured directly it is uncertain what types of circulations contribute most to the transport. There are three generic types of circulations that are candidates: wind-driven currents, thermohaline circulations, and mid-ocean eddies.

7.8.1 Wind-Driven Currents

The warm western boundary currents such as the Gulf Stream and the Kuroshio and their associated mid-ocean drift currents play an important role in meridional energy transport in the oceans. The swift, warm currents that flow poleward along the western boundaries of the Atlantic and Pacific oceans are capable of carrying large amounts of heat poleward. The equatorward flow of relatively cold water in eastern boundary currents also contributes to the poleward energy flux. In addition to these horizontal gyre circulations, the subtropical oceans have a shallow vertical overturning in the top 700 m that contributes about as much poleward heat transport as the gyre circulations (Bryden et al., 1991). The trade winds drive a poleward drift of warm water at the surface, which is balanced by an equatorward flow of slightly colder water below the surface.

We can estimate the poleward heat flux associated with these circulations by considering the product of the mass flux of water and the temperature difference between the poleward and equatorward flowing at the same latitude. The mass flow is the velocity of the current times its area and the water density. From Fig. 7.11, we estimate that the Gulf Stream is 100 km wide and 600 m deep and has an average current of 0.5 m s^{-1}. Assuming a density of 10^3 kg m^{-3}, we can obtain an estimate of the mass flux in the Gulf Stream of 3.0×10^{10} kg s^{-1}, or 30 sverdrups.

$$\begin{aligned}\text{Density} \times \text{width} \times \text{depth} \times \text{speed} &= 10^3\ \text{kg m}^{-3} \times 100\ \text{km} \times 600\ \text{m} \times 0.5\ \text{m s}^{-1} \\ &= 3 \times 10^{10}\ \text{kg s}^{-1}\end{aligned}$$

This estimate of 30 sverdrups agrees with more detailed calculations of the flow through the Florida Straits (Bryden and Hall, 1980). The mass flux in the Gulf Stream increases significantly northward from Florida, as it incorporates more flow from the gyre. If we assume that the Kuroshio has a similar mass flux, then the total poleward mass flux in Northern Hemisphere western boundary currents is 6×10^{10} kg s^{-1}. To calculate the energy flux associated with this mass transport, we need the heat capacity of water (4281 J K^{-1} kg^{-1}) and the temperature difference between the poleward-flowing boundary currents and the equatorward-flowing water at the same latitude. We do not know this temperature difference precisely, and it is very dependent on whether the equatorward flow is above or below the thermocline. It is interesting to consider how big this temperature difference must be in order for the western boundary currents to produce a meridional heat transport of about half the maximum oceanic flux displayed in Fig. 2.15. To obtain an oceanic flux of 1.6 PW requires a temperature difference between the poleward-flowing and equatorward-flowing water of about 6°C.

$$\begin{aligned} c_w \rho_w v_w \,\text{area}_w\, \Delta T &= 4218\,\mathrm{J\,K^{-1}kg^{-1}} \times 6 \times 10^{10}\,\mathrm{kg\,s^{-1}} \times 6.3\,\mathrm{K} \\ &= 1.6\,\mathrm{PW} \end{aligned} \tag{7.18}$$

From Fig. 7.14, we estimate the average surface temperature gradient across the Atlantic at 25°N to be 4–6°C. If the equatorward return flow is primarily in the eastern Pacific or in interior below the thermocline, then it would be easy to produce the required oceanic heat flux with the western boundary currents and an associated colder interior return flow. Similarly, if we look at the vertical structure of the temperature for the subtropical North Atlantic in Fig. 1.11, the temperature drops rapidly by about 10°C from the surface to 500 m depth, so a shallow overturning circulation with poleward flow near the surface and a return flow below will transport a significant amount of heat.

7.8.2 The Deep Thermohaline Circulation

The mass flow of the deep thermohaline circulation is governed by the rate at which deep water can be formed at high latitudes. In the Northern Hemisphere, deep water is formed only in the Atlantic at high latitudes, and the formation rate is quite slow, since it takes several centuries to replace the deep water in the Atlantic. It is estimated that the average rate of deep-water formation in the North Atlantic is 1.5–2 × 10^{10} kg s^{-1} and in the Antarctic Ocean about 1 × 10^{10} kg s^{-1}. Since the mass flux of the deep thermohaline circulation is more than a factor of ten smaller than that associated with the western boundary currents, it is safe to assume that it has a much smaller influence on the total northward ocean heat flux than the wind-driven gyres, although it may have a significant effect in the far North Atlantic.

7.8.3 Mid-Ocean Eddies

The Gulf Stream and the Kuroshio spin off long-lived eddies via baroclinic and barotropic instabilities (Fig. 7.12). These are the oceanic analogs to the eddies that produce most of the atmospheric meridional energy transport in mid-latitudes. However, the role of eddies for heat transport in the subtropical ocean is likely much less than in the atmosphere, because of their smaller spatial scales compared with the scale of the oceans. Moreover, the oceanic eddies are best developed well poleward of the latitude of the maximum oceanic transport. The wind-driven and thermohaline circulations are likely to provide much more important contributions to the meridional heat flux in the subtropics. However, in the high latitude Southern Ocean, ocean eddies are very important in determining the overturning circulation and the response of the ocean to changed wind stress.

EXERCISES

1. Use the data in Figs 7.1 and 7.2 to estimate how much the salinity of the surface water of the Arctic Ocean would need to increase before the surface density would equal the potential density at 1000-m depth. How does this compare with the average salinity of the ocean?
2. What depth of seawater would need to freeze in the Arctic Ocean to produce the increase in salinity of problem 1 in the top 100 m of water? Assume that all salt is rejected from sea ice and enters the 100-m layer.
3. With the negative wind stress curl characteristic of today's climate, the wind-driven meridional flow (7.16) is an equatorward drift, which we can hypothesize occurs mostly in the thermocline or above it. How would the net heat transport produced by this drift and its return flow be different if, rather than a warm western boundary jet, the return flow were a slow poleward drift near the bottom of the ocean?
4. Discuss the ways in which the extension of the warm, saline Gulf Stream into the Norwegian and Labrador seas assists in the formation of dense water that can sink to the depths of the Atlantic Ocean.
5. Use Fig. 7.1 to estimate the initial and final density values of a kilogram of water that starts in the tropics with a temperature of 28°C and a salinity of 35‰ and flows on the surface in the Gulf Stream to the Norwegian Sea, where it arrives with a temperature of –1°C. Assume the water conserves its salinity en route and loses heat by sensible heat transfer.
6. As an alternative to problem 5, assume that the kilogram of water starts in the tropics, but is cooled by evaporation along its route rather than by sensible heat loss. Estimate the mass of water that is lost by evaporation en route, if the parcel arrives in the Norwegian Sea with a temperature of –1°C. Calculate the salinity on arrival, assuming that no horizontal mixing or precipitation occurs, and the salinity is well mixed through the top 100 m of the ocean. What is the density on arrival? When you compare the final density with that obtained in problem 5, is the effect of evaporation on the final density significant? Is it important to know the Bowen ratio for the parcel along its route?

7. Suppose that a wind stress is applied to the ocean, taking the following simple form.

$$\tau_x = \begin{cases} A\cos\left(\dfrac{\pi y}{L}\right) & -L < y < L \\ A & |y| > L \end{cases}$$

Derive an equation for the vertical velocity at the bottom of the Ekman layer assuming a constant Coriolis parameter $f = f_o = 2\,\Omega \sin 30^\circ$. Derive an equation for the integrated meridional transport V_I, using (7.14) with the f and β appropriate for 30°N latitude. Determine a numeric value for the maximum w_E and V_I using the following constants: $A = 2$ dyn cm^{-2} = 0.2 N m^{-2}, L = 1500 km, ρ_o = 1025 kg m^{-3}. Plot τ_x, w_E, and V_I on the interval $-L < y < L$. Assuming that the ocean basin is 5000 km wide, calculate the water mass flux at 30°N associated with the interior flow. Compare this number with the estimate for the Gulf Stream mass flux given in Section 7.8.