8: OCEAN CIRCULATION

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Critical Concepts

- **CC1 Density and Layering in Fluids**: Water in the oceans is arranged in layers according to the water density. Many movements of water masses in the oceans, especially the movements of deep water masses, are driven by differences in water density.
- **CC3 Convection and Convection Cells**: Fluids, including ocean water, that are cooled from above, sink when their density is increased. This establishes convection processes that are a primary cause of vertical movements and the mixing of ocean waters. These processes are also important in transporting and distributing heat and carbon dioxide between the atmosphere and oceans and between regions of the globe.
- **CC6 Salinity, Temperature, Pressure, and Water Density**: Sea water density is controlled by temperature, salinity, and, to a lesser, extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Movements of water below the ocean surface layer are driven primarily by density differences.
- **CC7 Radioactivity and Age Dating**: Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope. Radioactive isotopes, especially those that were released during the period of atmospheric testing of nuclear weapons, are useful as tracers that can reveal the movements of water masses in the ocean, especially the rates at which deep water masses are formed.
- **CC9 The Global Greenhouse Effect**: The oceans and atmosphere are both important in studies of the greenhouse effect, as heat and carbon dioxide and other greenhouse gases are exchanged between the atmosphere and oceans at the sea surface. The oceans store large amounts of heat and carbon dioxide both in solution and in carbonates.
- **CC10 Modeling**: The complex interactions between the oceans and atmosphere can best be studied by using mathematical models. The motions of water masses within the body of the oceans, especially motions below the surface layer, are also studied extensively using mathematical models because they are extremely difficult to observe directly.
- **CC11 Chaos**: The nonlinear nature of ocean-atmosphere interactions makes at least parts of this system behave in sometimes unpredictable ways. It also makes it possible for changes in ocean circulation to occur in rapid, unpredictable jumps between one set of conditions and a different set of conditions, and these changes can affect climate.
- **CC12 The Coriolis Effect**: Water and air masses move freely over the Earth and ocean surface while objects on the Earth's surface, including the solid Earth itself, are constrained to move with the Earth in its rotation. This causes moving water or air masses to appear to follow curving paths across the Earth's surface. The apparent deflection, called the Coriolis effect, is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The deflection is at a maximum at the poles, is reduced at lower latitudes, and becomes zero at the equator.
- **CC13 Geostrophic Flow**: Air and water masses flowing on horizontal pressure gradients are deflected by the Coriolis effect until they flow across the gradient such that the pressure gradient force and Coriolis effect are balanced, a condition called geostrophic flow. This causes ocean currents and winds to flow around high and low pressure regions (regions of elevated or depressed sea surface height) in near circular paths. The circular gyres that dominate the global circulation of ocean waters are the result of water masses flowing geostrophically.

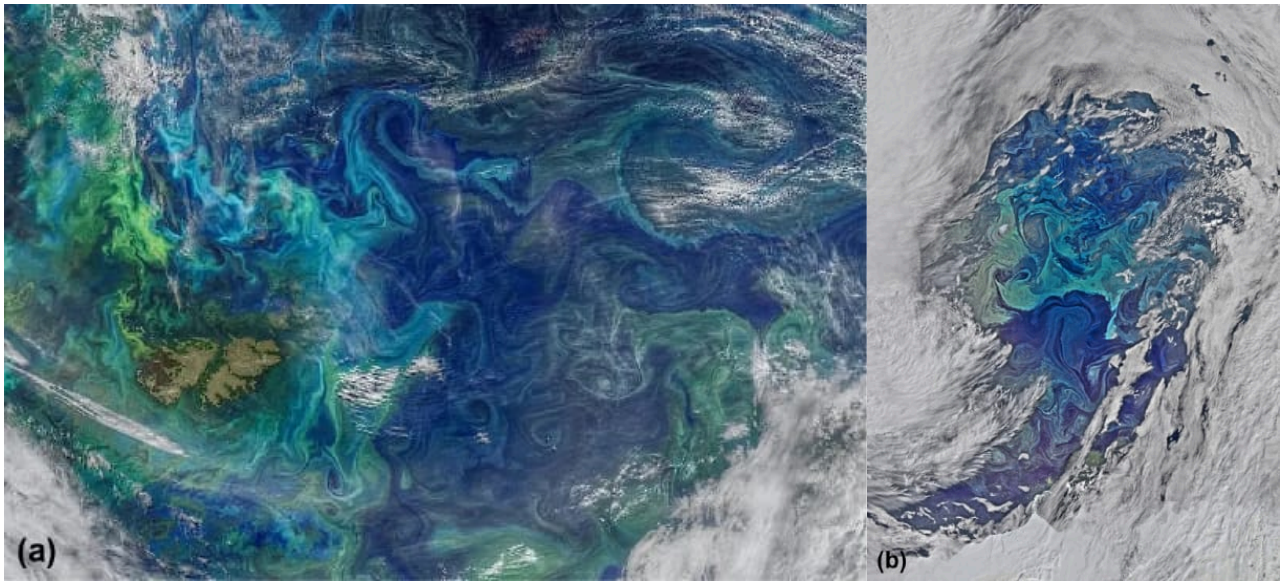

Figure 8-1. These images are scanned from a satellite with a visible/infrared radiometer that senses changes in radiated energy at multiple wavelengths. The data is converted to a false color image that shows elevated chlorophyll concentrations (a proxy for phytoplankton concentration) as the lighter colored blue-green swirls. The radiometers can not see through clouds so most of the bright white parts of the images are cloud cover. The complex swirls and patterns revealed in all these images are caused by eddies and other surface water motions, some of which come and go within a few hours. The eddy motions are too small and too short lived to be observed by oceanographers using ship-based measurements, so their existence was not known until observations were made from space. Eddies exist in all parts of the oceans. (a) South Atlantic Ocean near the Falkland Islands (visible on the left of the image). (b) The Southern Ocean just to the west of Drake Passage (the passage between the tip of South America and the tip of the Antarctic Peninsula).

Throughout the oceans, from surface to seafloor, water is in continuous motion, ranging in scale from movements of individual molecules to oceanwide movements of water by tidal forces and winds. Oceanographers generally group ocean-water movements into three categories: waves, **tides**, and ocean circulation. Waves (**Chap. 9**) and tides (**Chap. 10**) are primarily short-term (lasting from seconds to hours) oscillating motions that move water in orbital paths around a single location.

Seawater properties such as **salinity**, temperature, and chemical concentrations vary with location and depth. However, large volumes of water have nearly uniform properties in many areas and these are called water masses. Water masses are transported across the oceans by **currents**, and vertically up or down by **convection** (**CC3**) or by interaction with seafloor topography. As this circulation occurs, water masses mix with other water masses at their interface (usually a horizontal interface between layers; **CC1**). Mixing is the exchange of water between adjacent water masses, which results in a new water mass with a proportional combination of the properties of the individual water masses before mixing. Currents and mixing are the major processes that transport and distribute heat, dissolved chemicals, **suspended sediments**, and planktonic bacteria, algae, and animals in the oceans.

Chapter 7 describes how ocean surface currents can affect local climates. This chapter examines the physical processes that control present-day ocean circulation, reviews the major features of circulation, and describes some aspects of the relationship between ocean circulation and global climate, both past and present. Subsequent chapters describe how ocean circulation controls and/or affects the distribution and abundance of life in the oceans. Although waves and tides have relatively little influence on large-scale ocean circulation, they contribute substantially to local currents, particularly in coastal waters and **estuaries**, and to vertical mixing, particularly in the upper several hundred meters of the water column (**Chaps. 9, 10**).

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8.1: Energy Sources

Winds are the primary energy source for currents that flow horizontally in the ocean surface layers (less than 100 to 200 m deep). Hence, surface currents are often called “wind-driven currents” or “wind drift currents.” Currents that flow deep in the oceans below the level affected by winds are generated primarily by convection caused by variations in water **density** (**Chap. 5**, **CC3**). Higher-density waters sink and displace (push aside or up) less-dense deeper water. Because density is determined by temperature and salinity (**CC6**), deep-water circulation is called **thermohaline circulation**.

The primary energy source for both wind-driven circulation and thermohaline circulation is the sun. Winds are generated by density variations in the Earth’s atmosphere caused by solar heating and radiative and evaporative cooling (**Chap. 7**). Thermohaline circulation is driven by density differences between ocean water masses caused by temperature and salinity variations. Variations in temperature and salinity are controlled, in turn, primarily by balances between solar heating and radiative cooling, and between solar evaporation and precipitation, respectively (**Chap. 7**).

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8.2: Wind-Driven Currents

On a windy day at the **shore**, we can easily see that winds cause ocean water motions such as waves. Winds also create currents that can transport large volumes of water across the oceans.

Generation of Currents

When winds blow across the ocean surface, energy is transferred from wind to surface water as a result of the **friction** between the wind and the surface. The energy transferred to the ocean surface sets the surface layer of water in motion, generating both waves and currents. The process of energy transfer from winds to waves and currents is complex and depends on many factors, including wind speed, **surface tension** of the water, and roughness of the surface (that is, whether waves are already present, their height, and whether they are breaking). Therefore, the percentage of wind energy that is converted into **kinetic energy** of ocean currents is highly variable. With a steady wind, the speed of the surface current is generally between 1% and 3% of the wind speed, so a wind of 60 $km \bullet h^{-1}$ will generate a surface current of about 1 to 2 $km \bullet h^{-1}$.

Surface water set in motion by the wind flows horizontally across the water below it. Because of internal friction between the surface water and the water below, wind energy is transferred downward. If we consider the water column to consist of a series of very thin horizontal layers, we can envision the moving surface layer transferring some of its kinetic energy to the next lower layer by friction, setting that layer in motion. The second layer, in turn, transfers some of its resulting kinetic energy to the layer below it, setting that layer in motion (Fig. 8-2a). However, only a fraction of the kinetic energy of each moving layer is transferred to the layer below it. Consequently, the speed of a wind-driven current decreases progressively with depth. Wind-driven currents are restricted to the upper 100–200 m of the oceans and generally to even shallower depths.

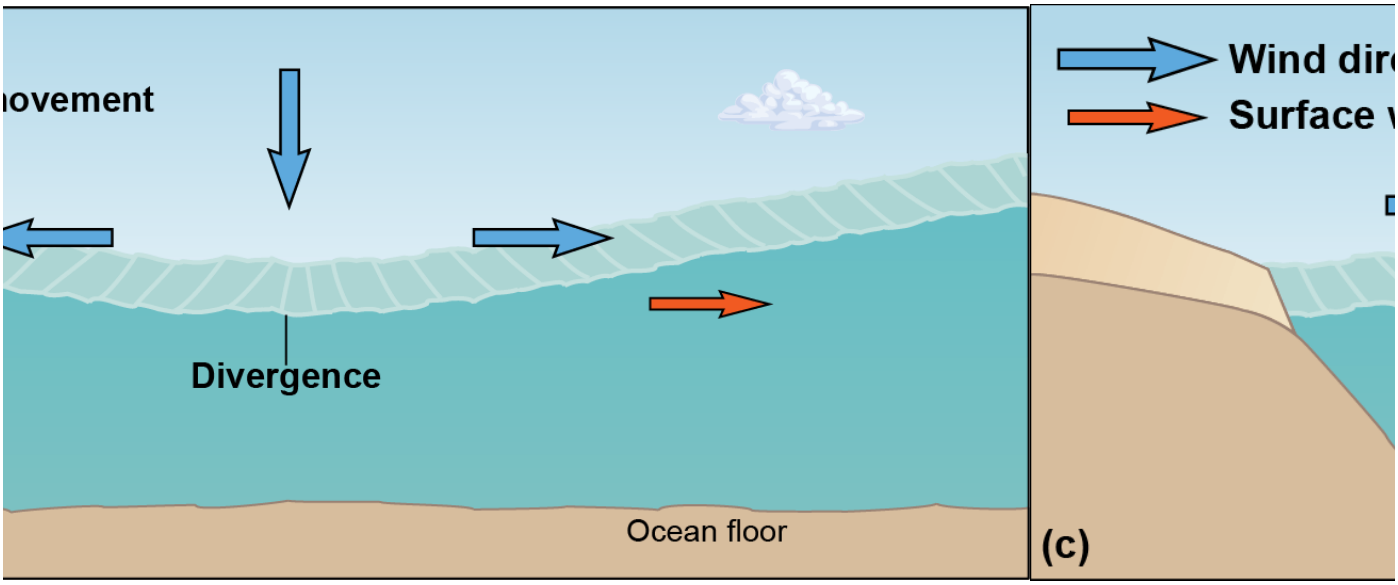

Figure 8-2. (a) Wind energy is converted to water movements, called currents, by friction between the moving air and the water surface. The resulting kinetic energy of the water at the surface is transferred vertically downward into the water column by friction between the water molecules. Friction occurs when the molecules of one layer are moved, as shown in the enlarged diagram of three adjacent molecules. As the upper-layer molecules are moved, the distance between them and the molecules immediately below changes. At point A, the distance between molecules is slightly decreased. With a decrease in distance the repulsive force between the molecules' electron clouds increases faster than the attractive gravitational force between the molecules (partly because the electron clouds are closer to each other than the two centers of mass are). Thus, the molecule below is "pushed" forward and downward. Conversely, the distance between molecules at the point labeled B is slightly increased. Since the repulsive force between the electron clouds is reduced by more than the gravitational force is reduced, the molecule below is "pulled" forward and upward. This process transfers kinetic energy downward through successive layers of molecules. (b) Sloped sea surfaces can be created in the open ocean when winds transport the surface water layer either away from a divergence, as shown here, or toward a convergence that then becomes an elevated area of sea surface. (c) Sloped sea surfaces can also be created at a coastline when winds transport the surface water layer offshore or onshore. The sea surface slopes are greatly exaggerated in these diagrams.

Restoring Forces and Steering Forces

Once current motion has been started, it will continue for some time after the wind stops blowing because the water has momentum. It is like a bicycle that continues to roll forward after the rider stops pedaling. Just as a bicycle slows and eventually stops because of friction in the wheel bearings and between tires and road, ocean currents slow and would eventually stop if winds stopped blowing and did not restart. The energy of a current is dissipated by friction between water layers flowing over each other. However, the frictional force between moving layers of water is small. This is one reason why the currents created by winds can flow for long periods after the winds stop and can also flow into and through regions with little wind, such as the Doldrums (**Figs. 7-10**, **7-11**).

There is a second reason for the continued flow of wind-created currents after the winds stop. Transport of surface water by wind-driven currents while winds are blowing can cause the sea surface to be sloped (**Fig. 8-2b,c**). Sea surface slopes created by wind-driven transport are extremely small, no more than a few centimeters of height difference across hundreds or thousands of kilometers of ocean surface. However, when the surface is sloped, a horizontal **pressure gradient** is formed (**CC13**), aligned in the same direction as the sea surface slope. The water flows from high pressure toward low pressure and tends to restore the ocean surface to a flat horizontal plane.

Once set in motion, the speed and direction of any current are modified by friction and three other factors. First, any body in motion on the Earth, including water moving in currents, is subject to deflection by the **Coriolis effect** (**CC12**). Second, the presence of coasts can block current flow and cause the water to mound up or to be deflected. Third, current speed and direction are affected by the presence of horizontal pressure gradients.

The following sections examine how the interactions between climatic winds and the restoring and steering forces account for the location and characteristics of surface currents in the oceans (**Fig. 8-3**).

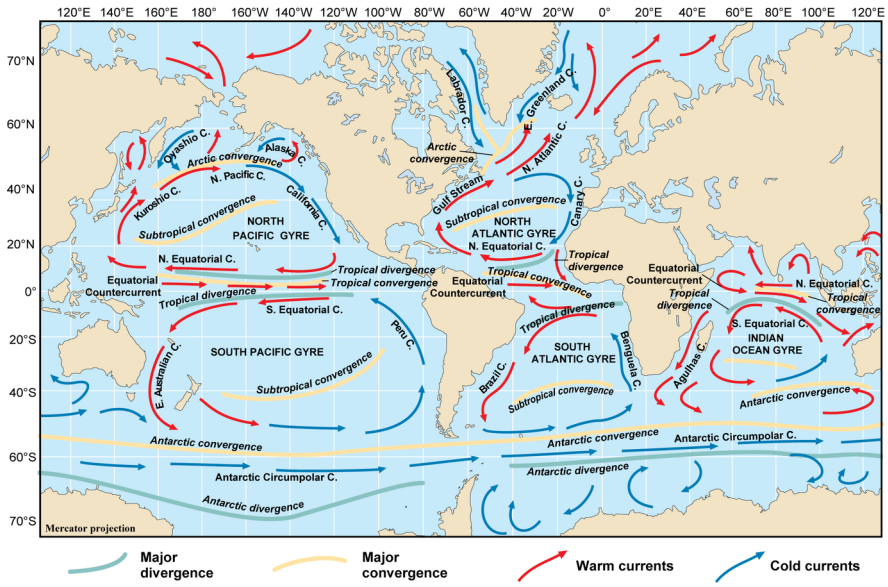

Figure 8-3. The wind-driven surface currents of the oceans and the locations of the major convergences and divergences are related. The surface currents form subtropical gyres both north and south of the equator in each ocean basin, and there are subtropical convergences within these gyres. Convergences are also found at higher latitudes between the subtropical gyres and the Antarctic Circumpolar Current, and between the subtropical gyres and the complicated high-latitude gyres in the Northern Hemisphere. Divergences lie on either side of the equator between the subtropical gyres and surrounding Antarctica.

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8.3: Ekman Motion

During his Arctic expedition in the 1890s, Fridtjof Nansen noticed that drifting ice moved in a direction about 20° to 40° to the right of the wind direction. The deflection occurs because water set in motion by the wind is subject to the Coriolis effect (**CC12**). The surface current that carries drifting ice does not flow in the direction of the wind but is deflected ***cum sole*** ("with the sun"), in the same way we see the sun move across the sky—to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (**CC12**). Remember the term *cum sole*. It always means "to the right in the Northern Hemisphere and to the left in the Southern Hemisphere."

The Ekman Spiral

As surface water is set in motion by the wind, it successively sets in motion a series of thin layers of water (**Fig. 8-2a**), transferring energy down into the water column. The surface layer is driven by the wind and deflected by the Coriolis effect. Each successively lower layer is driven by the layer above, with speed decreasing in each lower layer, and each layer is also deflected by the Coriolis effect. When it is set in motion by friction with the layer above, each layer's direction of motion is deflected by the Coriolis effect *cum sole* to the direction of the overlying layer. This deflection establishes the Ekman spiral (**Fig. 8-4**), named for the physicist who developed the mathematical relationships to explain Nansen's observations that floating ice does not follow the wind direction.

Surface water set in motion by the wind is deflected at an angle *cum sole* to the wind (**Fig. 8-4**). Ekman's theory predicts this angle to be 45° *cum sole* to the wind direction under ideal conditions, with constant winds and a water column of uniform density. Under normal conditions, the deflection is usually less. As depth increases, current speed is reduced, and the deflection increases. The Ekman spiral can extend to depths of 100 to 200 m, below which the energy transferred downward from layer to layer becomes insufficient to set the water in motion. This depth is a little greater than the depth where the flow is opposite to the surface current direction. This depth is called the "depth of frictional influence." At this depth, the current speed is about 4% of the surface current speed. The water column above the depth of frictional influence is known as the "wind-driven layer."

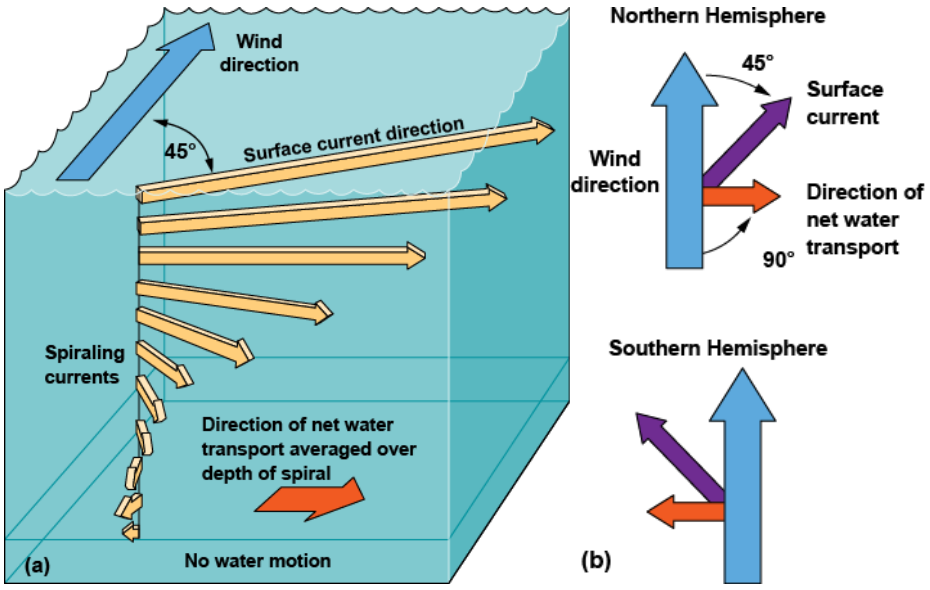

Figure 8-4. The Ekman spiral. (a) In the Northern Hemisphere, the Coriolis deflection causes a surface current to flow at an angle to the right of the wind direction. As the kinetic energy of the wind-driven current is transferred downward in the water column, each layer of water is deflected to the right of the direction of the layer above, producing a spiraling current called the Ekman spiral, in which speed decreases with depth. (b) If the wind blows long enough over deep water, the Ekman spiral motion becomes fully developed. When the spiral is fully developed in the Northern Hemisphere, the surface current direction is 45° to the right of the wind direction, and the net water transport is 90° to the right of the wind direction. In the Southern Hemisphere, the deflection and net water transport are to the left.

The most important feature of the Ekman spiral is that water in the wind-driven layer is transported at an angle *cum sole* to the wind direction and that the deflection increases with depth. Ekman showed that, under ideal conditions, the mean movement

of water summed over all depths within the wind-driven layer is at 90° *cum sole* to the wind. Such movement is called **Ekman transport**.

For the Ekman spiral to be fully established, the water column above the depth of frictional influence must be uniform or nearly uniform in density, water depth must be greater than the depth of frictional influence, and winds must blow at a constant speed for as long as a day or two. Because these conditions are rare, the Ekman spiral is seldom fully established.

Pycnocline and Seafloor Interruption of the Ekman Spiral

In most parts of the ocean, there is a range of depths within which density changes rapidly with depth (Fig. 8-5). Such vertical density gradients are called pycnoclines (**CC1**). In most parts of the open ocean, a permanent pycnocline is present with an upper boundary that is usually at a depth between 100 and 500 m. In some mid-latitude areas, another shallower seasonal pycnocline develops in summer at depths of approximately 10 to 20 m. Where a steep (large density change over a small depth increment) pycnocline occurs within the depth range of the Ekman spiral, it inhibits the downward transfer of energy and momentum. The reason is that water layers of different densities slide over each other with less friction than do water layers of nearly identical density. Water masses above and below a density interface are said to be frictionally decoupled. When a shallow pycnocline is present, the wind-driven layer is restricted to depths less than the pycnocline depth, and the Ekman spiral cannot be fully developed. Pycnoclines and the vertical structure of the ocean water column are discussed in more detail later in this chapter.

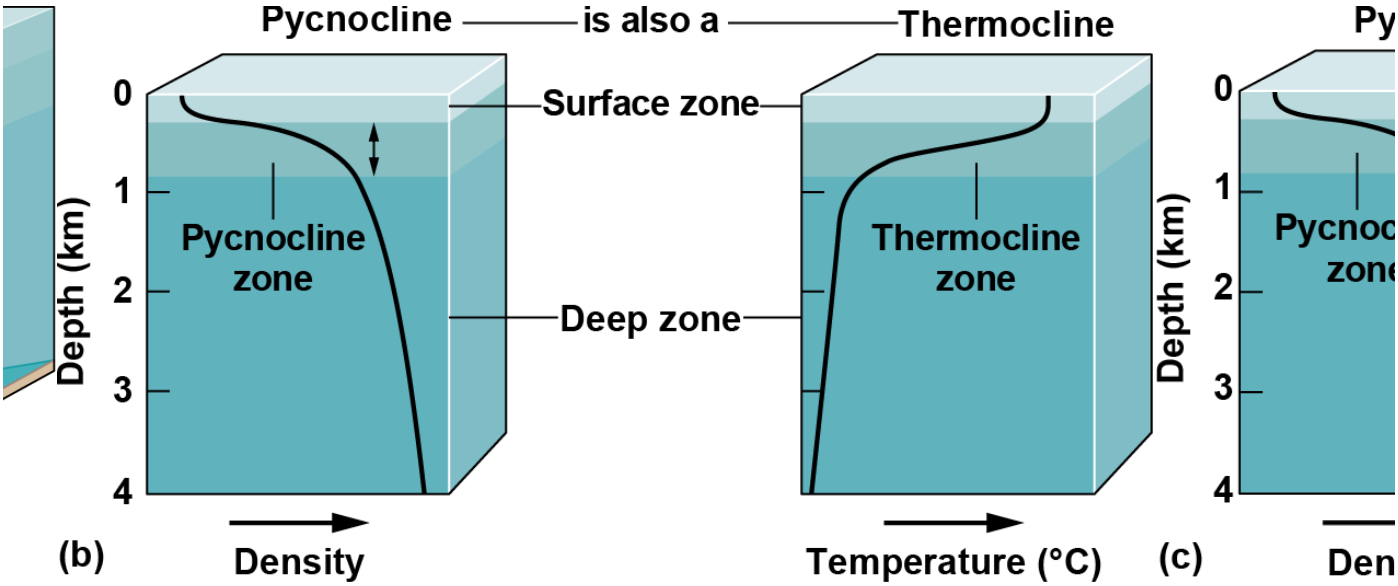

Figure 8-5. Density stratification in the oceans. (a) Throughout most of the open oceans (except in high latitudes where surface waters are cooled), a surface or mixed layer of relatively warm, nutrient-depleted, low-density water overlies a layer called the "pycnocline zone" in which density increases rapidly with depth. Below the pycnocline zone is a deep zone of cold, nutrient-rich water where density increases more slowly with depth. (b) The rapid density change with depth in the pycnocline is due primarily to changes in temperature, and the pycnocline is often called a "thermocline." Shallow seasonal thermoclines can also develop above the main pycnocline and in coastal waters because of solar heating of the surface water. (c) In some other regions, and particularly in shallow water near major freshwater inputs, a pycnocline may develop in which the density change with depth is due to vertical variations in salinity. In this case, the pycnocline is also a "halocline."

The Ekman spiral also cannot be fully established in shallow coastal waters because of added friction between the near-bottom current and the seafloor. Where the Ekman spiral is not fully developed because of a shallow pycnocline or seafloor, the surface current deflection is somewhat less than 45°, and Ekman transport is at an angle of less than 90° to the wind. In water only a few meters deep or less, the wind-driven water transport is not deflected from the wind direction. Instead, the current generally flows in the wind direction, but it is steered by bottom **topography** and flows along the depth **contours**.

Where Ekman transport pushes surface water toward a coast, the water surface is elevated at the **coastline** (Fig. 8-6a). If the Ekman transport is offshore, the sea surface is lowered at the coastline. Similarly, where Ekman transport brings surface water together from different directions, a **convergence** is formed at which the sea surface becomes elevated (Fig. 8-6b). In contrast, where winds transport surface water away in different directions, a **divergence** is formed at which the sea surface is depressed. These elevations or depressions are exaggerated in the diagrams.

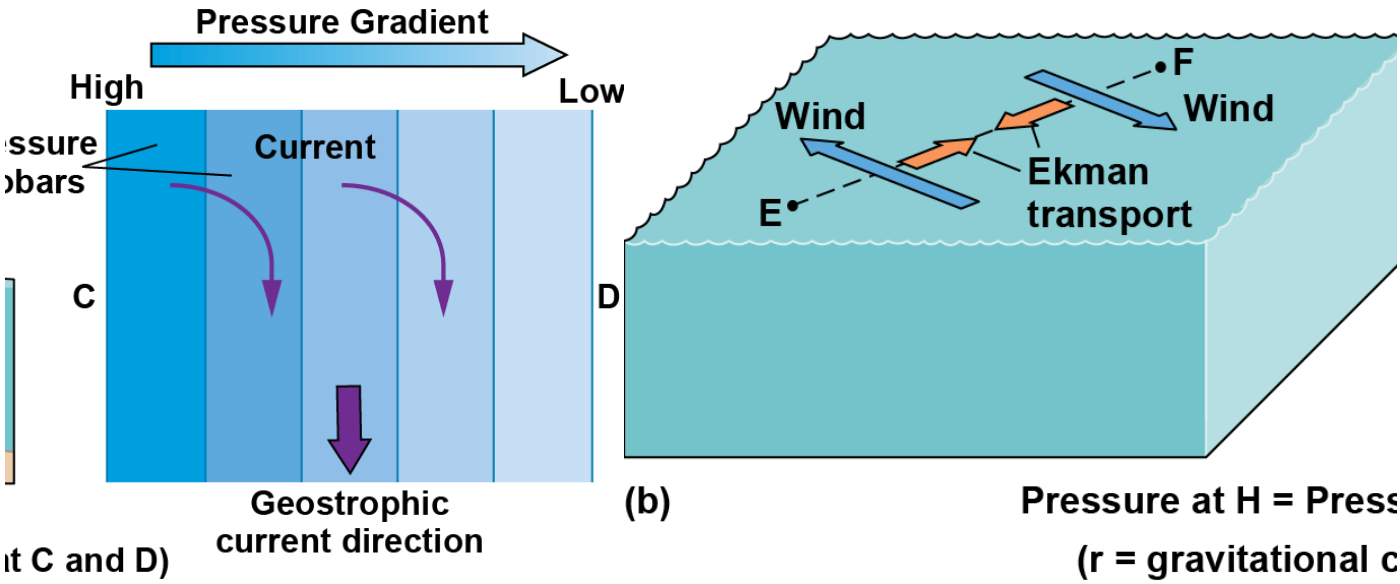

Figure 8-6. Ekman transport, geostrophic currents, and Ekman circulation. (a) Where winds blow surface water shoreward, a horizontal pressure gradient develops that causes water below the surface layer to flow offshore, and water to be downwelled near the coastline. This flow is deflected by the Coriolis effect such that, if the system reaches equilibrium, the lower-layer geostrophic current flows parallel to the coast (across the pressure gradient). (b) Away from the coastline, in locations where Ekman transport of surface waters tends to create a mound of surface water at a convergence, a horizontal pressure gradient develops that causes geostrophic currents below the surface layer to flow across the pressure gradient, and water downwells at the convergence. The circulation caused by the Ekman transport, downwelling, and geostrophic flow in the lower layer is called “Ekman circulation.” Ekman circulation occurs in reverse at divergences or where winds move surface water offshore.

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8.4: Geostrophic Currents

Ekman transport of surface layer water tends to produce sloping sea surfaces by “piling up” the water in some locations and “removing” surface water from others. **Geostrophic** currents are the result of horizontal pressure gradients in the water column below the surface caused by such variations in sea surface level (**CC13**).

Ekman Transport, Sea Surface Slope, and Pressure Gradient

If winds blew continuously, Ekman transport would cause the sea surface slope to increase continuously unless other water movements caused the sea surface to return to a flat configuration. Although sloping sea surfaces do develop, the slopes are always very small—so small that we perceive the sea surface to be flat even in regions of the most persistent winds. Water movements associated with Ekman transport that tend to restore the flat sea surface configuration are driven by the horizontal pressure gradients created under sloping sea surfaces (**Fig. 8-6**). Horizontal pressure gradients develop in response to the Earth’s gravitational force on the water.

The pressure gradient force tends to move water from the high-pressure region under the elevated part of the sea surface toward the lower-pressure region under the depressed part (**Fig. 8-6**). This direction is opposite to that of the Ekman transport. The flow induced by the pressure gradient tends to reduce the gradient and sea surface slope, restoring a flat sea surface.

Ekman transport occurs only in the wind-driven layer, which usually occupies only the upper few tens of meters of the water column. The mean water flow in the wind-driven layer is toward the highest point of sea surface elevation. However, the pressure gradient created when Ekman transport generates a sloping sea surface extends to depths beyond the wind-driven layer and is not restricted by shallow pycnoclines. As a result, the mean flow below the wind-driven layer is directed away from the area of maximum sea surface elevation (**Fig. 8-6**). In the area of maximum elevation, water is **downwelled** from the surface to below the wind-driven layer to join the pressure gradient flow.

Balance between Pressure Gradient and Coriolis Effect

All fluids, including air and water, are subject to the Coriolis effect when they flow horizontally (**CC12**). In the wind-driven layer, the Coriolis effect causes Ekman transport at approximately 90° *cum sole* to the wind direction. Ekman transport stores wind energy as **potential energy** associated with the elevation of the sea surface in one area relative to another. Water flows in response to the horizontal pressure gradient created by the sea surface elevation. As the water flows, it is accelerated and deflected *cum sole*. If allowed to develop long enough, the resulting currents, called “geostrophic currents” (**CC13**), are deflected to flow parallel to the contours of constant pressure (or along the side of the “hill” parallel to the sea surface height contours), and the Coriolis deflection is balanced by the pressure gradient force.

Geostrophic currents generated under ideal conditions beneath sloping sea surfaces are shown in **Figure 8-6**. These ideal conditions would be established only if winds blew at uniform speed over the entire area represented in the figure for several days or more. Such equilibrium conditions are never fully met in the oceans. Consequently, **Figure 8-6** shows only the general characteristics of geostrophic currents. In reality, the currents are continuously changing as winds vary.

Geostrophic current speed and direction are determined by the magnitude and distribution of the horizontal pressure gradient (**CC13**). The pressure gradient can be determined from the sea surface slope and the water density. For example, the pressure at point C in **Figure 8-6a** is greater than the pressure at the same depth at point D by an amount that is equal to the additional weight per unit area of water column at point C. The additional weight is equal to the water density multiplied by the difference in sea surface height between the locations. Using these simple relationships, oceanographers have been able to estimate some surface current speeds by measuring sea surface height differences very precisely with satellite radar sensors.

Sea surface slopes that are created by wind-driven currents and that sustain geostrophic currents are extremely small. For example, one of the fastest ocean currents is the Florida Current, a part of the Gulf Stream that flows around the tip of Florida from the Gulf of Mexico into the Atlantic Ocean. The Florida Current flows at about 150 $cm \cdot s^{-1}$ (5.4 $km \cdot h^{-1}$). This current flows on a sea surface with an approximate height difference of only 20 cm across about 200 km—a slope of 1 in 1 million. Most sea surface slopes are even smaller, and the geostrophic currents are correspondingly slower.

Dynamic Topography

Measuring very tiny sea surface height differences to determine current speeds would appear to be a very difficult task. However, oceanographers have learned to measure the differences indirectly. They carefully measure the density of seawater throughout the depth of the water column at different locations and calculate the **dynamic height**.

The dynamic height is the height of the water column calculated from a measured density profile of the water column between the surface and a depth below which it is assumed there are no currents. We can see how the calculation of dynamic height works if we remember that density is mass per unit volume. Consider several columns of water, each with a 1-cm^2 surface area, in different locations (**Fig. 8-7**). If the total mass of water in each column were the same, the height of the column with a lower average density would be greater than that of the column with higher average density (**Fig. 8-7**). From several calculated dynamic heights, we can estimate the average sea surface slope between locations. By making such calculations for many different locations, we can draw a detailed map of sea surface height called a "dynamic topography map." Detailed water density distribution data can also be used to identify slopes on pycnoclines, which are then used to estimate geostrophic current speeds and directions in deep-ocean water masses.

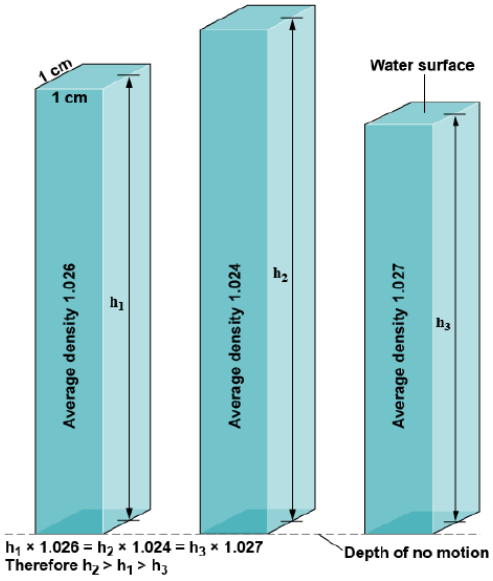

Figure 8-7. Calculations of dynamic height can be used to determine sea surface slopes. The height of the water column above a "depth of no motion" is calculated from measurements of the distribution of density with depth. The depth of no motion is a depth below which it can be reasonably assumed that there are no currents flowing. If the average density of the water column above the depth of no motion is known for several locations, the height of the sea surface at each location and the horizontal pressure gradients between the locations can be calculated.

This procedure for calculating dynamic height or sea surface slope is relatively simple, but it involves assumptions that are usually not fully met. First, we must select a depth of no motion below which we assume there are no currents. The pressure at this depth must be the same at all locations; otherwise, the horizontal pressure difference would cause water to flow horizontally. Because pressure does not vary at the depth of no motion, the mass of water above this depth must be the same at any location.

Until just the past few years, dynamic topography maps were generated from measurements made at many different individual locations at various times, sometimes months or years apart. The necessary measurements of salinity and temperature were made from research vessels, which take many weeks to complete a series of measurements at stations spaced across an ocean. As a result, until relatively recently, dynamic topography maps represented average conditions over many weeks or months, and we knew very little about short-term (up to months) variations in geostrophic currents.

Oceanographers still rely on indirect calculations of dynamic height to investigate geostrophic currents. However, the development of unattended moored arrays, autonomous floats, and satellite sensors now allows simultaneous data to be gathered at many locations. Moored vertical strings (arrays) of instruments at key locations and autonomous floats that profile the water column can collect and send back to shore-based facilities continuous data that can be used to calculate dynamic height. More than 4000 autonomous floats are currently deployed. Satellite radar sensors have now become sensitive enough to measure sea surface heights directly, and with sufficient precision that the data can be used to draw detailed dynamic topography maps of any part of the ocean every few days. Although the satellite data give only the dynamic height at the sea surface, these data,

when combined with the data collected by moored and autonomous instruments, have drastically improved our ability to observe geostrophic currents. As a result, we are just beginning to understand their complexities and variability.

Energy Storage

Geostrophic currents flowing near the surface are generated by winds because they depend on Ekman transport to establish the sea surface slopes under which they flow. However, geostrophic currents continue to flow even if winds abate, because Ekman transport stores wind energy as potential energy by piling up the water. The stored potential energy is converted to momentum to maintain geostrophic currents during intervals when winds are light or cease.

The length of time during which geostrophic currents continue to flow after winds abate depends on the magnitude of the sea surface slope and the area over which the slope occurs. The larger the area of sloping sea surface and the greater the slope, the longer it takes for geostrophic currents to restore the sea surface after winds abate. Where local winds create sea surface slopes over small areas, geostrophic currents restore the sea surface and cease to flow within a few hours after the winds abate. This is often the case for geostrophic currents created by the interaction of storms and coastal water masses.

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8.5: Open-Ocean Surface Currents

Surface currents of the open oceans (Fig. 8-3) are wind-driven currents initiated by Ekman transport and maintained as geostrophic currents. Although they typically extend to depths of several hundreds of meters, they may extend as deep as 2000 m in some limited cases. The most obvious features of ocean surface currents are the **gyres** in subtropical **latitudes** of each ocean. Separate **subtropical gyres** are present north and south of the equator in each ocean except in the Indian Ocean, where the Northern Hemisphere gyre would occur at a location occupied by landmasses. Subtropical gyres in the Northern Hemisphere flow clockwise; those in the Southern Hemisphere flow counterclockwise.

At latitudes above the subtropics, ocean surface currents are more complicated. The North Atlantic and North Pacific oceans have high-latitude gyres similar to, but less well formed than, the subtropical gyres. The high-latitude gyres rotate in the opposite direction (counterclockwise) from the northern subtropical gyres. In the Southern Hemisphere, high-latitude gyres are absent because the surface current that flows clockwise from west to east is not interrupted by a landmass, so it flows completely around Antarctica.

Geostrophic Currents on a Water-Covered Earth

If there were no continents, geostrophic surface layer currents would be relatively simple. The climatic winds and **convection cells** that would be present on a rotating, water-covered Earth are shown in Figure 7-10. Figure 8-8a shows these idealized climatic winds, and the directions in which the wind-driven layer of ocean water would be moved by Ekman transport due to the winds. In the **trade wind** and polar easterly zones, Ekman transport is partly toward the pole and partly toward the west. In the **westerly** wind zones, Ekman transport is toward the equator and to the east. Surface water layer divergences and **upwelling** of subsurface water occur at the equator and at the atmospheric upwelling regions between the polar and Ferrel cells in each hemisphere. Surface water convergences occur at the atmospheric downwelling regions between the Hadley and Ferrel cells in each hemisphere. Thus, Ekman transport would produce a depression of the sea surface at the equator and between the polar and Ferrel cells in each hemisphere, and an elevation of the sea surface between the Hadley and Ferrel cells in each hemisphere (Fig. 8-8b).

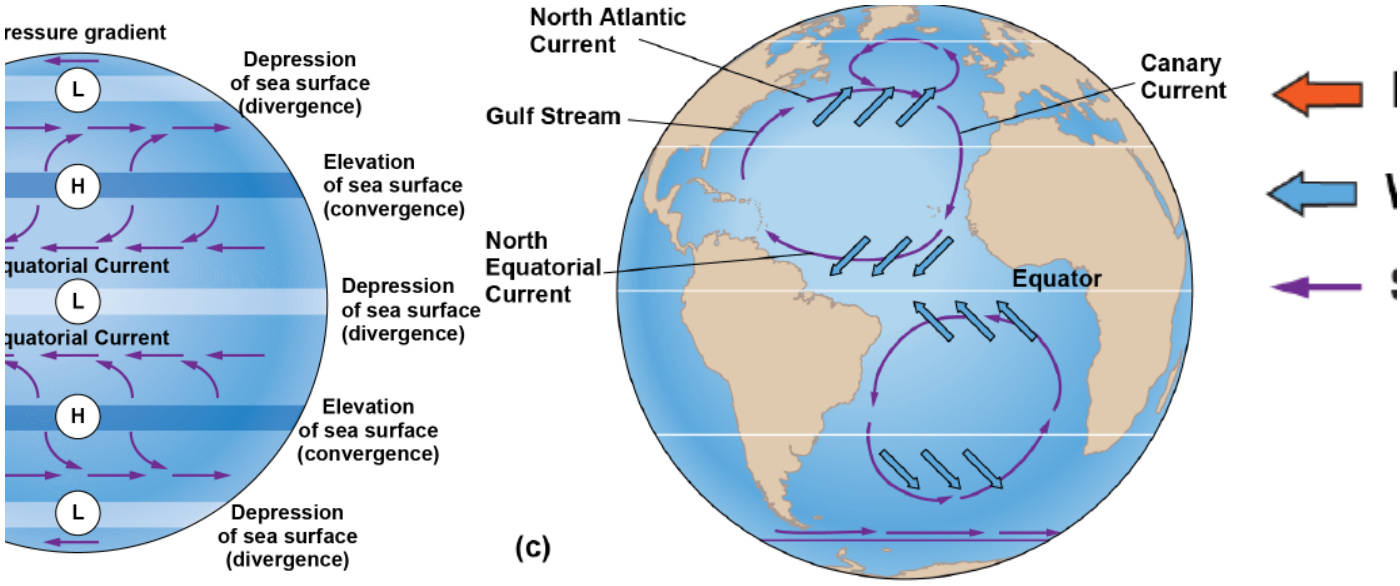

Figure 8-8. Ocean surface layer currents are driven by climatic winds and modified by the presence of landmasses. (a) The climatic winds on an idealized water-covered Earth would cause Ekman transport of surface waters in the directions shown. (b) Those water movements would lead to the formation of convergence and divergence, and geostrophic currents would flow on the pressure gradients created by the Ekman transport of the surface layer. These geostrophic currents would flow around the Earth from east to west, or vice versa, between the latitudes of the convergences and divergences. (c) However, landmasses block the flow of geostrophic currents and divert them to the north or south, resulting in the formation of geostrophic gyres.

The sloping sea surfaces created by climatic winds establish horizontal pressure gradients within the water column and thus

establish geostrophic currents. If an equilibrium were reached, the geostrophic currents would flow around the Earth from east to west or from west to east across the pressure gradients (**Fig. 8-8b**). Currents would flow to the west in the trade wind zones and polar regions and to the east in the westerly wind zones. Geostrophic currents on the real Earth cannot flow in this way because they are interrupted by continents in all but the westerly wind zone around Antarctica. The Antarctic Circumpolar Current does, in fact, flow from west to east around the Earth in the Southern Hemisphere westerly wind zone (**Fig. 8-3**), just as the no-continent model suggests (**Fig. 8-8b**).

Geostrophic Gyres

Around each of the three major oceans, continents provide a western boundaries (the Atlantic coast of North and South America, Pacific coast of Asia and Australia, and Indian Ocean coast of Africa) and eastern boundaries (the Atlantic coast of Europe and Africa, Pacific coast of North and South America, and Indian Ocean coast of Southeast Asia and Australia). Within each ocean, geostrophic east-to-west or west-to-east currents are established at the same latitudes and in the same locations as those in the simple ocean-covered Earth model (**Figs. 8-3, 8-8b**). Where these currents meet a continent, they are blocked and diverted to the north or south.

The westward-flowing currents within the trade wind zones of the Northern and Southern Hemispheres are called the North Equatorial Current and the South Equatorial Current, respectively (**Fig. 8-3**). These Equatorial currents carry large volumes of water from east to west until they meet the western boundary continent. Surface water “piles up” there, and the Equatorial currents are deflected away from the equator. The currents then flow toward higher latitudes as **western boundary currents** until they enter the westerly wind zone, where they join the west-to-east geostrophic flow. When the currents reach the eastern boundary land mass, they are deflected primarily toward the equator. After turning toward the equator, they flow as **eastern boundary currents** until they enter the trade wind zone, rejoin the Equatorial Current, and complete their circuit, or gyre (**Figs. 8-3, 8-8c**).

As western boundary currents flow toward the pole, they are subject to the Coriolis effect. Therefore, if there were no opposing horizontal pressure gradient, they would be deflected to the east. Similarly, as eastern boundary currents flowed toward the equator, they would be deflected to the west (**Fig. 8-9a**). In each case, the deflection would be toward the center of the gyre. In addition, Ekman transport of surface water in the trade wind and westerly wind zones and the Coriolis deflection of the gyre currents, including the boundary currents, all tend to move water toward the center of the gyre. As a result, an equilibrium situation is created in which there is a mounded sea surface at the center of the gyre (**Fig. 8-9b**). Currents that make up the gyre flow around the mound in geostrophic balance between the Coriolis effect and the horizontal pressure gradient that surrounds this central location.

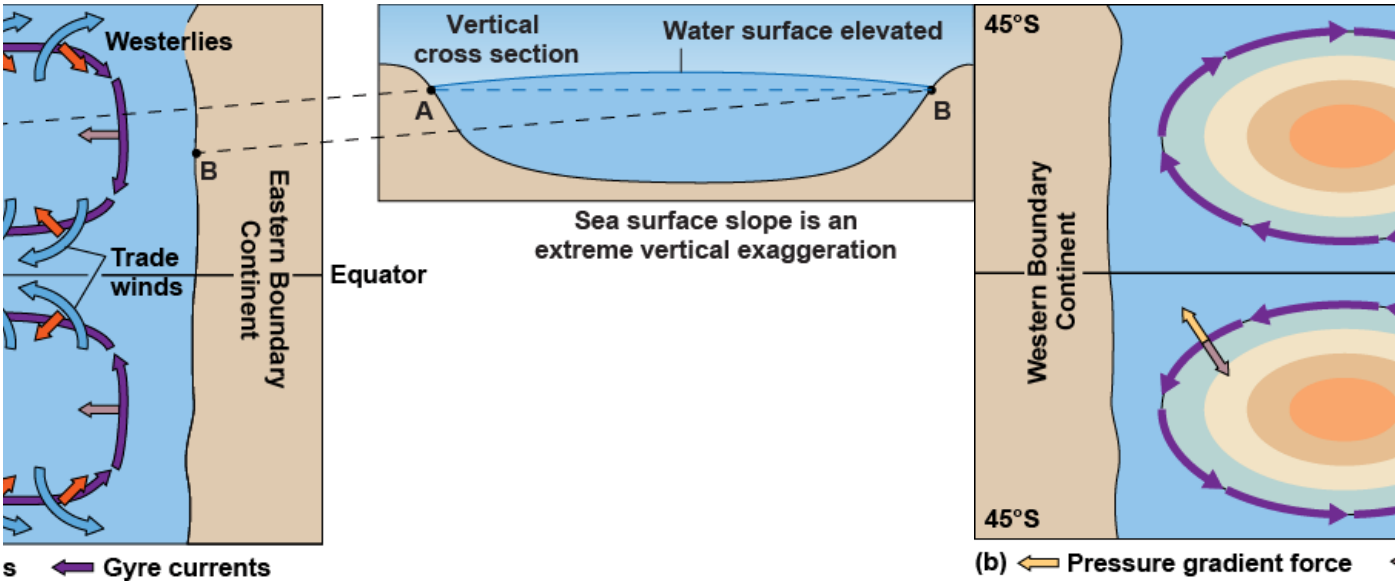

Figure 8-9. Geostrophic currents of the subtropical gyres. (a) Wind-driven Ekman transport due to the trade winds and westerlies, and the Coriolis deflection of the boundary currents, both act to elevate the ocean surface at the center of each subtropical gyre. (b) This elevated sea surface creates a horizontal pressure gradient that tends to move water outward. As water flows on this pressure gradient, it is deflected by the Coriolis effect and flows geostrophically around the mound. At equilibrium, the sea surface elevation does not vary because the tendency of water to be moved to the center of the mound is balanced by the pressure gradient, which generates geostrophic currents. Thus, the wind energy of the trade winds and westerlies is transferred into the geostrophic currents that flow continuously and form the subtropical gyres.

Hence, gyres are geostrophic currents that flow on mounded surfaces. The center of each subtropical gyre is called a "subtropical convergence" (**Fig. 8-3**). Subtropical gyres are essentially the same as the idealized gyres shown in **Figure 8-9**. However, they are modified in shape and location by the configuration of each ocean's coastlines (**Fig. 8-3**).

Geostrophic ocean gyres act like giant flywheels, which use angular momentum to store rotational energy. They spin at an almost constant speed that represents the average wind energy input to the gyre. If winds cease or are abnormally light, the geostrophic current gyre, or flywheel, slows very gradually because turbulence and friction between the gyre currents and the adjacent water masses or seafloor dissipate the energy very slowly. When winds resume, Ekman transport builds the height of the mound, increasing the sea surface slope and speed of the gyre. This process occurs very slowly because massive volumes of water must be moved to elevate the sea surface even a millimeter or two over a mound that is hundreds or thousands of kilometers in diameter. Winds that drive the ocean gyres are variable over periods of days, but very long periods of calm or unusually high winds are rare. Because gyre current speeds change very slowly, they are relatively stable and generally do not reflect normal wind speed variations.

Westward Intensification of Boundary Currents

Western boundary currents include the Gulf Stream and the Brazil, Kuroshio, East Australian, and Agulhas Currents (**Fig. 8-3**). Eastern boundary currents include the California, Peru, Canary, and Benguela Currents. Western boundary currents are narrower, faster, deeper, and warmer than eastern boundary currents (**Table 8-1**). They are generally so deep that they are constrained against the edge of the **continental shelf** and do not extend across the continental shelf to the shore.

Table 8-1. *Boundary Currents*

Properties	Western Boundary Currents	Eastern Boundary Currents
Northern Hemisphere examples	Gulf Stream, Kuroshio Current	California Current, Canary Current
Southern Hemisphere examples	Agulhas Current, Brazil Current	Peru Current, Benguela Current
Width	Narrow (≤100 km)	Broad (≈1000 km)
Depth	Deep (to 2 km)	Shallow (≤500 m)
Speed	Fast (>100 km•day^{-1})	Slow (<50 km•day^{-1})
Volume transport	Large (50×10^{6} m^{3}•s^{-1})	Small (10–15×10^{6} m^{3}•s^{-1})
Boundaries with coastal currents	Sharply defined	Diffuse
Upwelling	Almost none	Frequent

Properties	Western Boundary Currents	Eastern Boundary Currents
Nutrients	Depleted	Enhanced by upwelling
Fishery	Usually poor	Usually good
Water temperature	Warm	Cool

The reasons for the westward intensification of boundary currents are quite complex. However, the major reason is related to the Earth's rotation. Western boundary currents are intensified because the strength of the Coriolis effect varies with latitude.

We can look at westward intensification in a simplified way. The Coriolis effect, which increases from the equator to the poles, is weaker in the latitudes of the trade wind band than in the westerly wind band. Therefore, water moving eastward is deflected more quickly toward the south in the westerly wind zone than water moving westward is deflected toward the north in the trade wind zone (**Fig. 8-10**). Consequently, in the westerly wind zone, geostrophic flow tends to transport surface water toward the center of the gyre over the entire width of the ocean. In contrast, surface water transported westward in the trade wind zone tends to flow with less deflection (that is, in a circle of larger radius; **CC12**) across the ocean (**Fig. 8-10**), is constrained at the equator by the Ekman transport of the trade winds in the other hemisphere, and tends to pile up on the west side of the ocean.

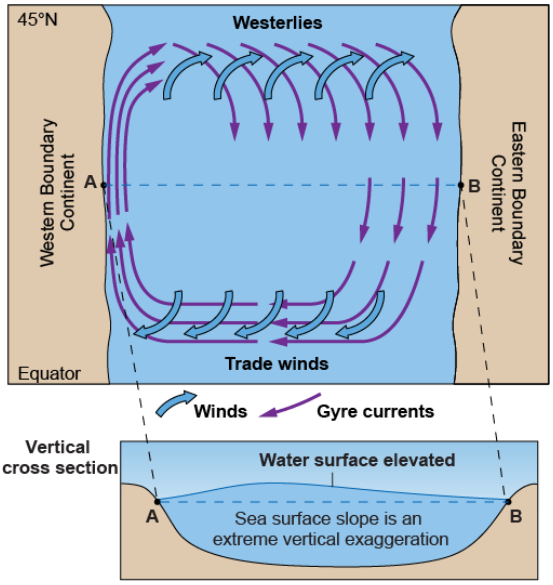

Figure 8-10. The westward-moving geostrophic currents of the subtropical gyres flow near the equator, where the Coriolis deflection is weak, whereas the eastward-moving currents flow where the Coriolis deflection is much greater. Hence, the westward-moving water near the equator flows across the ocean more directly, piling up water on the west side. The greater Coriolis deflection further from the equator tends to turn parts of the eastward-moving gyre currents toward the equator even before it reaches the eastern boundary continent, "spreading" the southerly part of the gyre currents that returns water from the westerly region toward the equator.

As a result, the mound at the center of the subtropical gyre is offset toward the west side of the ocean (**Fig. 8-11**). The western boundary current is laterally compressed against the continent, and the sea surface slope is steeper toward the western boundary than toward the eastern boundary. Steeper sea surface slopes cause faster geostrophic currents. Therefore, western boundary currents are faster than eastern boundary currents. The westward offset of the subtropical gyre mound causes the pycnocline to be depressed (deeper) on the west side of the ocean in relation to the east side (**Fig. 8-11**). Hence, western boundary currents are narrower and deeper than eastern boundary currents.

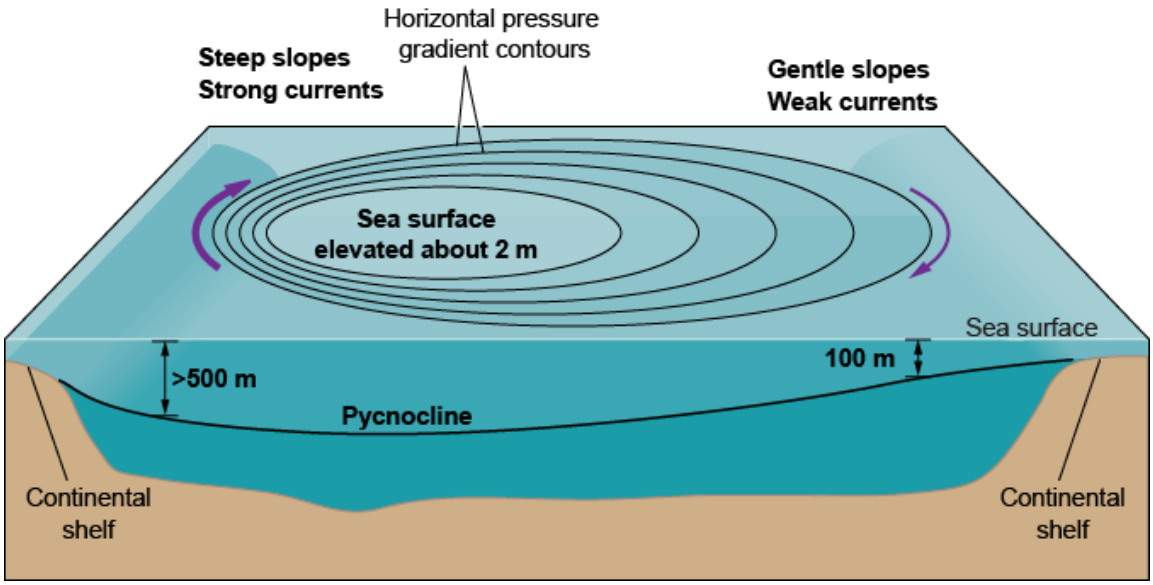

Figure 8-11. In the Northern Hemisphere, the sea surface elevation in the center of the subtropical gyres is displaced to the west. Thus, the horizontal pressure gradients are stronger and the gyre currents are faster on the west side of the ocean than on the east side. In addition, because of this displacement, the mixed layer is thicker on the west side and the deep layer does not extend onto the continental shelf.

This simplified explanation serves reasonably well. However, for a more detailed understanding of westward intensification, it is necessary to examine a difficult concept called “vorticity” (see **Online Box 8B1**).

The characteristic differences between western and eastern boundary currents have major consequences for processes that sustain fisheries off the adjacent coasts (**Chap. 12**). The differences are also an important factor in the transport of heat poleward (**Chap. 7**). Warm western boundary currents are fast-flowing and deep. They have little time to cool and only a small surface area from which to lose heat as they move into higher latitudes. Therefore, heat energy carried away from the equator by western boundary currents is released to the atmosphere primarily when the water enters the westerly wind zone, where the gyre currents flow eastward. Heat transported by western boundary currents moderates the climates of regions into which they flow. For example, the Gulf Stream transports heat north and then east across the Atlantic Ocean, moderating western Europe’s climate.

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8.6: Equatorial Surface Currents

The Northern and Southern Hemisphere trade wind zones are separated by the Doldrums, where winds are very light and variable (**Figs. 7-10, 7-11**). In the simplified model of subtropical gyre formation discussed in the preceding section, the Northern and Southern Hemisphere trade winds produce Ekman transport of surface water to the northwest and southwest, respectively (**Fig. 8-8**). This pattern would be present if the Doldrums (the atmospheric **intertropical convergence zone**) were exactly over the equator. However, atmospheric circulation interacts in a complex manner with continental landmasses, which are concentrated in the Northern Hemisphere. One result of this interaction is that the Doldrums are displaced somewhat to the north of the equator during most of the year (**Fig. 7-11**).

The displacement of the Doldrums leads to a complex system of west to east and east to west currents that flow on or very close to the equator. These currents and their behavior are important components in the **El Niño** Southern Oscillation (ENSO) described in **Chapter 7**. The most important characteristic of Equatorial Currents is that they are able to flow directly east to west or west to east across entire oceans without being deflected by the Coriolis effect since the Coriolis deflection is zero at the equator and negligibly small within a few degrees north and south of the equator. Equatorial Currents are discussed in more detail in **Online Box 8B2**.

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8.7: High-Latitude Surface Currents

In the North Atlantic and North Pacific, secondary gyres occur at higher latitudes than the subtropical gyres. They are called “subpolar gyres” and rotate in the direction opposite that of the adjacent subtropical gyres. Subpolar gyres are wind-driven in the same way as subtropical gyres. However, the prevailing winds of the region, the polar easterlies, are much more variable than trade winds.

Because of this wind variability and the complex shapes of the ocean basins in polar regions, the subpolar gyres are more complex and variable than the subtropical gyres.

In the Southern Hemisphere, subpolar surface currents are different from their Northern Hemisphere counterparts. The primary reason is the lack of continents in the Southern Hemisphere. Because of the wide connection between the oceans, Southern Hemisphere subpolar currents flow all the way around the Antarctic continent (**Fig. 8-3**). At high latitudes, a geostrophic current, the East Wind Drift, flows westward around Antarctica. It flows in response to the pressure gradient caused by the elevation of the sea surface near the Antarctic continent due to southerly Ekman transport in the polar easterly wind belt. Farther from Antarctica, an eastward current, the Antarctic Circumpolar Current (or West Wind Drift), flows geostrophically around Antarctica under a sea surface that slopes up to the north as a result of the Ekman transport of the westerlies (**Fig. 8-3**). The Antarctic Circumpolar Current is combined with, and a part of, the eastward-moving currents of the Southern Hemisphere subtropical gyres in each ocean.

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8.8: Upwelling and Downwelling

Although wind-driven surface water motions are primarily horizontal, winds can also cause vertical water movements. Vertical movements occur because wind-driven Ekman transport drives a layer of surface water up to about 100 m deep across the underlying water layers. Winds moving surface water away from an area cause a divergence at which the surface water is replaced by water upwelling from below (**Fig. 8-12a**). Upwelling is important because colder water upwelled from below the pycnocline has high concentrations of **nutrients**, such as nitrogen and phosphorus compounds. The nutrients are needed to fertilize **phytoplankton**, the microscopic algae and **phototrophic** bacteria and archaea that grow only in near-surface waters and that are the principal source of food supporting all animal life in the oceans. Most surface waters are deficient in nutrients and cannot sustain phytoplankton growth unless nutrients are supplied by upwelling or recycling (**Chap. 12**).

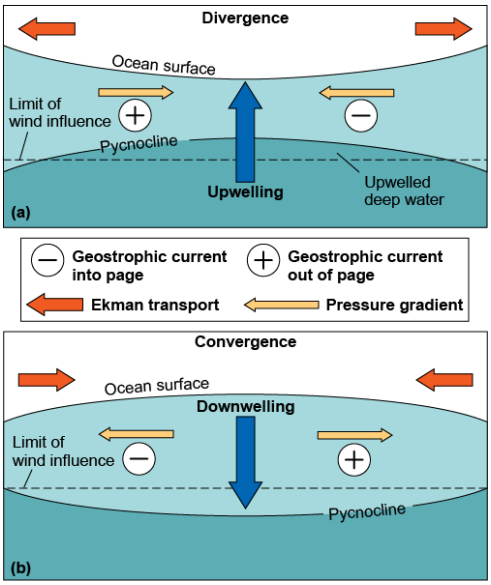

Figure 8-12. Divergences and convergences are areas of upwelling and downwelling in the open oceans. (a) At a divergence, the surface layer is thinned and the pycnocline depth reduced. If wind mixing extends below the pycnocline depth, cold, nutrient-rich, deep water is upwelled and mixed into the surface layer. (b) At a convergence, the surface layer is thickened, and the pycnocline is driven deeper. If the pycnocline is below the maximum depth of wind influence, there is no mixing of cold, nutrient-rich subpycnocline water into surface layers, even when winds are very strong.

At a convergence, winds moving surface waters toward an area elevate the sea surface and create a high-pressure region in the subsurface water, causing downwelling (**Fig. 8-12b**). Surface water is transported downward at the convergence and then outward below the wind-driven layer under the influence of the horizontal pressure gradient. Downwelling thickens the layer of warmer surface water above the pycnocline and tends to isolate cold subpycnocline water, preventing it from mixing upward into the surface layers. Downwelling also occurs in areas where the surface water density is increased by cooling or evaporation sufficiently that this surface water has a higher density than the water below.

Locations of Upwelling and Downwelling

Divergences are upwelling areas where primary productivity is high because of the supply of nutrients from below. Convergences are downwelling areas where productivity is poor. The principal open-ocean divergences include a band that parallels the equator between the North and South Equatorial Currents and a band around Antarctica between the Antarctic Circumpolar Current and the eastward coastal flow around the continent (**Fig. 8-3**). In the equatorial region, upwelling is inhibited by the flow of warm surface water from west to east in the Equatorial Countercurrents (**Online Box 8B2**). Consequently, persistent upwelling of nutrient-rich, cold subpycnocline water at the equatorial divergence is limited primarily to the east side of the Pacific Ocean off the Peruvian coast. Upwelling in this region is inhibited by the movement of warm surface water from west to east during El Niño (**Fig. 7-18b**). The principal open-ocean convergences are in the center of each of the subtropical surface current gyres (**Fig. 8-3**).

Coastal upwelling can occur when winds blow parallel to the coast and cause Ekman transport of the surface layer offshore.

Surface water transported offshore is replaced by deeper water that moves inshore and is upwelled (**Fig. 8-13a**). **Coastal downwelling** occurs when surface water is transported toward shore displaces near-shore surface water downward and forces the deeper water to move offshore (**Fig. 8-13b**). These processes are particularly important to local ecosystems in areas where there is a shallow pycnocline, with warm, nutrient-depleted water at the surface and colder, nutrient-rich water below. In such areas, offshore transport causes coastal upwelling that supplies large quantities of nutrients to support phytoplankton growth (**Chap. 12**).

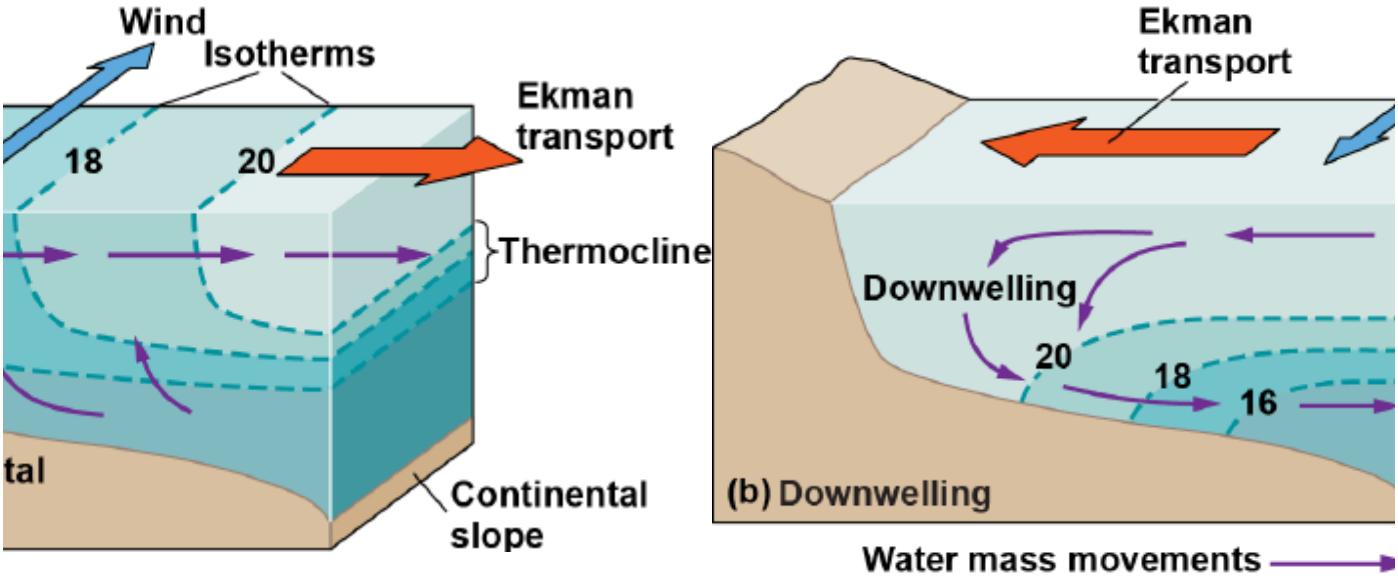

Figure 8-13. Coastal upwelling and downwelling. (a) When winds produce Ekman transport of the surface water layer in an offshore direction, the pycnocline (usually a thermocline) is raised, sometimes all the way to the surface, and cold, nutrient-rich, deep water is upwelled into the surface layer. (b) When winds produce Ekman transport of the surface water layer in an onshore direction, the pycnocline (again usually a thermocline) is depressed, shelf surface water is downwelled, and cold, nutrient-rich, deep water is forced deeper and off the shelf, so it cannot enter the surface layer, even if winds are strong enough to mix coastal water to the depth of the continental shelf break.

Boundary Currents and Upwelling or Downwelling

The characteristics of ocean gyre boundary currents strongly influence the occurrence and persistence of coastal upwelling near the basin's coasts. Western boundary currents are deep and generally compressed against the continental shelf edge. Hence, a deep layer of warm surface water generally lies over the cold, nutrient-rich subpycnocline water and prevents it from extending onto the continental shelf (**Fig. 8-13b**). Because coastal winds normally move aside a surface layer only a few tens of meters deep, offshore Ekman transport in western boundary current regions leads to upwelling of warm, nutrient-poor water from within the deep surface layer (**Fig. 8-13b**). Thus, coastal upwelling of nutrient-rich water is rare in western boundary current regions.

Eastern boundary currents are relatively shallow and wide, and they extend onto the continental shelf. Cold, nutrient-rich, deep water can migrate onto the shelf as a bottom layer below the shallow, warm surface layer (**Fig. 8-13a**). Only moderate coastal winds with offshore Ekman transport are needed to cause cold, nutrient-rich waters to upwell to the surface. Consequently, coastal upwelling is more frequent, widespread, and persistent on the eastern boundaries of the oceans (west coasts of continents) than on the western boundaries of the oceans (east coasts of continents).

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8.9: Coastal Currents

Coastal currents are independent of and more variable than the adjacent oceanic gyre boundary currents. The zone in which coastal currents flow varies in width along different coasts. For example, the coastal current zone is wide off the Mid-Atlantic coast of North America because the continental shelf is wide, and the western boundary current (Gulf Stream) flows in deeper water off the edge of the shelf. In contrast, the coastal current zone off the Pacific coast of North America is narrow because the continental shelf is narrow, and the relatively shallow eastern boundary current flows over the narrow shelf almost to the coastline. Variable winds, tides, freshwater inflow from rivers, friction with the seafloor, and steering by coastline irregularities, such as capes, affect currents more in coastal waters than in the open ocean.

Coastal currents generally flow parallel to the coastline in a direction determined by the winds (and by Ekman transport). Coastal winds are more variable in speed and direction than are trade winds and other global winds that drive the ocean gyres. Coastal currents can be established quickly in response to storm winds, but they can also disappear or change direction within a few hours.

The strongest coastal currents occur when strong winds blow in areas with large freshwater inputs from rivers. In such areas, a shallow, low-salinity layer of surface water is formed over a very steep pycnocline. The surface layer slides easily over layers below this pycnocline. Thus, the wind energy is concentrated in the shallow, surface-layer current, rather than being transmitted and distributed throughout a greater depth.

Because they are directed by local wind patterns and interaction with the coastline, coastal currents may flow in the opposite direction from the adjacent ocean gyre boundary currents. For example, coastal currents off the Atlantic coast of North America generally flow to the southwest from Labrador to Cape Hatteras and along the Carolina, Georgia, and Florida coasts (**Fig. 8-14**) in the direction opposite that of the adjacent Gulf Stream.

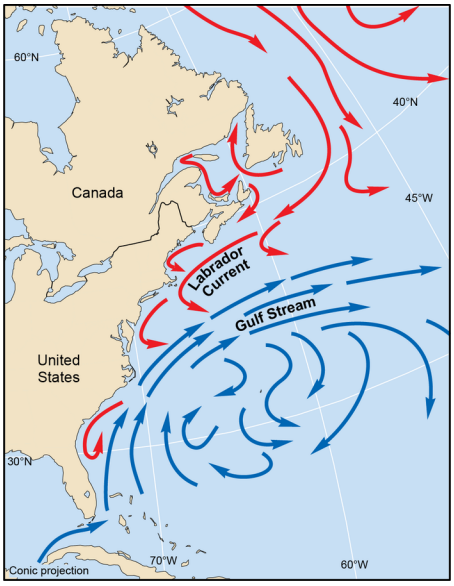

Figure 8-14. Off the Atlantic coast of North America, the coastal currents are cold-water currents that flow to the south on the continental shelf inshore from the northward-flowing Gulf Stream.

The boundary between the warm Gulf Stream and colder coastal water, called a **front**, can be clearly discerned in satellite imagery as a distinct difference in water color. The nearshore water is greenish or brownish, whereas Gulf Stream water is a clear, deep blue. The color difference is due to the greatly reduced concentrations of suspended sediment and phytoplankton in Gulf Stream water (**Chap. 5**).

Off the west coast of North America, the boundary between the water of the California Current (which is part of the North Pacific Ocean subtropical gyre) and the coastal water is not well defined. The main reason is that the California Current is a weak, diffuse eastern boundary current. During winter and spring, a weak and variable coastal current, the Davidson Current, flows north in the opposite direction from the adjacent boundary current. The Davidson Current is driven by the predominant winds of winter and

spring that blow from the southwest. These winds produce Ekman transport of water onshore and a geostrophic flow to the north on the resulting sea surface slope (**Fig. 8-15a**). In summer and fall, prevailing winds change to blow from the north. They cause Ekman transport offshore, a generally southward-flowing coastal current, and upwelling (**Figs. 8-13, 8-15b**). The summer and fall upwelling is responsible for the very high primary productivity and abundant sea life in coastal waters off the west coast of North America.

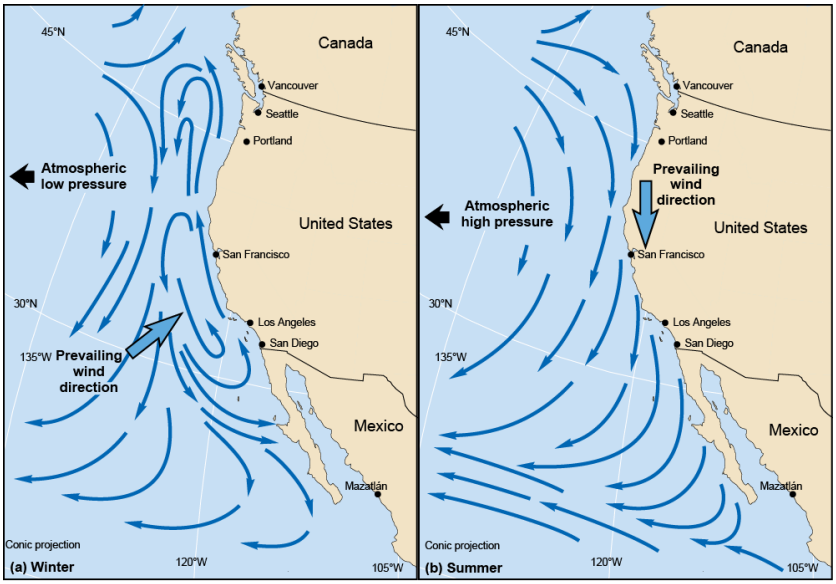

Figure 8-15. Coastal currents reverse seasonally off the Pacific coast of North America. (a) They flow to the north during winter and sometimes spring, when there is an area of atmospheric low pressure offshore and the coastal winds blow mainly from the southwest. (b) In the summer, when there is atmospheric high pressure offshore and coastal winds are mainly from the north, the coastal currents flow to the south, in the same direction as the offshore southward current that is part of the subtropical gyre in the North Pacific Ocean.

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8.10: Eddies

Chapter 7 describes how winds are arranged in global patterns that determine climate. It also explains that winds are highly variable in any given location. The local variations are part of **weather** systems, which we readily associate with the swirling patterns of clouds and precipitation. The global ocean current systems are analogous to the climatic winds, and the oceans have their own "weather systems," which includes variable swirling motions and meandering fronts, just like atmospheric weather. These swirling motions are called **eddies**.

The principal difference between atmospheric weather and ocean current "weather" is the scale of the eddy motions. Ocean currents generally move more slowly than most atmospheric winds. Consequently, the Coriolis effect tends to make ocean currents flow in curving paths that have a smaller radius than comparable atmospheric eddies (**CC12**). In other words, ocean eddies are generally much smaller than atmospheric eddies. Because ocean eddies are smaller, they are more numerous than atmospheric eddies.

Satellites can detect many atmospheric circulations and patterns that contribute to the weather. However, such **synoptic** measurements cannot be made so easily within the body of the oceans. Therefore, we do not have a detailed understanding of deep ocean current variability.

Satellite Observations of Eddies

Satellite infrared sensors detect ocean surface temperature by measuring the sea surface heat radiation. Satellite optical sensors tuned to different **wavelengths** of light can detect ocean color by measuring **back-scattered** sunlight, creating images that look similar to photographs. Such satellite observations are most effective for observing ocean surface currents where temperature and/or color differ markedly between water masses. Unfortunately, in most areas, the differences are very small. The easiest surface currents to observe by satellite are western boundary currents because the water they carry is warmer than the adjacent coastal waters. Areas near river discharges are also good for satellite observations of currents because the suspended particles in river outflow alter ocean color.

Even in areas where satellite observations are most effective, color or temperature variations of the sea surface are so small that they must be computer-enhanced to be seen by researchers. Computer enhancement of satellite images generally assigns bright false colors to areas that are slightly different in temperature and true color, thus making the demarcations between currents or water masses more prominent. Figure 8-16 is a color-enhanced infrared satellite image of the Gulf Stream region of the Atlantic Ocean.

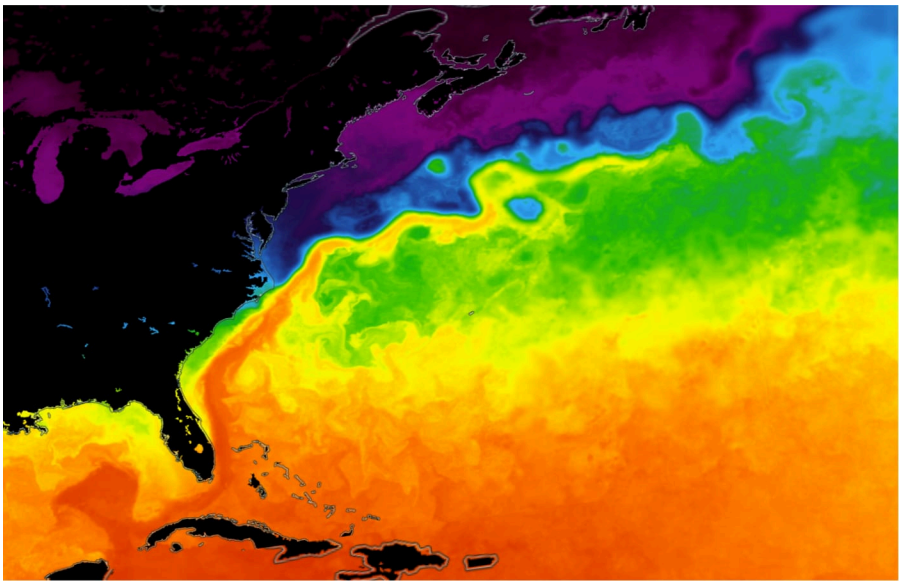

Figure 8-16. This image of the northwestern Atlantic Ocean shows the sea surface temperature derived from NOAA satellite sensors. Temperatures are depicted by the false colors in this image. Shades of red denote the areas of highest temperature, and progressively lower temperatures are depicted by orange, yellow, green, blue, and violet shades. The image shows the Gulf Stream separating the coastal water mass from the Sargasso Sea water mass. The complex meanders of the Gulf Stream and both cold-core and warm-core eddies are easy to see. For example, the isolated blue area surrounded by yellow and green in the offshore water is a cold core eddy, and the green areas isolated in the light blue coastal water are warm core eddies.

Mesoscale Eddies

The image in **Figure 8-16** shows the warm Gulf Stream flowing northward from the tip of Florida along the edge of the continental shelf. The front between this water and the colder coastal water is very sharp (**Fig 8-14**). A similar front can be seen between Gulf Stream water and colder Sargasso Sea water in the interior of the North Atlantic subtropical gyre. The infrared satellite image shows that there is great complexity along these fronts. We can see numerous secondary fronts and **meanders** of the Gulf Stream. We also see isolated, almost circular areas of warm water on the coastal-water side of the Gulf Stream and similar isolated areas of cold water on the Sargasso Sea side called warm-core and cold-core rings.

These Gulf Stream meanders and rings are eddies that continuously form and change shape and location. Meanders travel slowly northward along with the Gulf Stream and can become larger as they move. If a meander becomes tight enough, it can break off the Gulf Stream entirely and form a spinning ring of water. If the meander is to the coastal-water side of the Gulf Stream, a warm-core ring that spins clockwise is formed and isolated (**Fig. 8-17**). If the meander is to the Sargasso Sea side of the Gulf Stream, a cold-core ring that spins counterclockwise is formed as colder coastal water is pinched off inside the meander (**Fig. 8-17**).

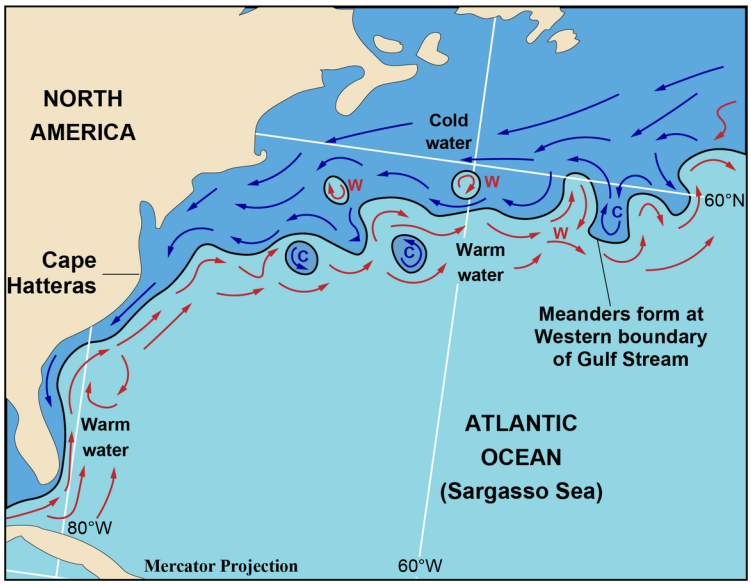

Figure 8-17. Gulf Stream rings. As the Gulf Stream passes north of Cape Hatteras and no longer flows along the edge of the continental shelf, meanders form. The meanders become more extreme as they move north and may eventually be pinched off to form warm- or cold-core rings. The rings generally drift south and dissipate or are reabsorbed by the Gulf Stream. Meander and ring formation is variable and complex, as shown in **Figure 8-16**.

The rings are from 100 to 300 km across and can extend to the seafloor. They are encircled by swiftly flowing currents that move at approximately 90 cm•s^{-1}. Both cold- and warm-core rings drift southeastward and slowly disappear as the core water temperature changes to match that of the surrounding water. Many rings are reattached to, and reabsorbed by, the Gulf Stream. Cold-core rings can maintain their integrity for up to several months depending on their direction of travel, but warm-core rings do not generally last as long, because of the shallower water and narrow shelf they travel over.

Gulf Stream rings transport heat, dissolved substances (including nutrients), and marine organisms that are weak swimmers or that prefer the warmer or colder water. Specific locations within the complex frontal system, with its associated warm- and cold-water rings, are better than others for certain fish **species**. Hence, the location of the Gulf Stream and its rings is important to fishers. This information is also important to ships if they are to minimize fuel costs by taking advantage of, or not fighting, ocean currents. The position of the Gulf Stream front and its rings is now monitored and forecasted. Many oceangoing fishing and other vessels obtain frequent up-to-date satellite images of the Gulf Stream region while at sea, just as they receive weather forecast maps.

Swiftly flowing meanders and rings are also present in other western boundary currents, such as the Kuroshio Current and eddies similar to Gulf Stream rings are present throughout the oceans. The general term for these eddies, including Gulf Stream Rings is “mesoscale eddies.” Most mesoscale eddies are less well defined than Gulf Stream rings, and their current speeds are generally lower. Mesoscale eddies, which in some cases extend all the way to the deep-sea floor, are the ocean equivalent of atmospheric high- and low-pressure zones. They range in diameter from 25 to 200 km, drift a few kilometers per day, and generally have rotating currents of between about 10 cm•s^{-1} and the 90 cm•s^{-1} of the Gulf Stream ring currents. In comparison, atmospheric depressions are about 1000 km across, travel about 1000 km per day, and sustain rotating winds of up to approximately 20 m•s^{-1}.

Mesoscale eddies transport and distribute heat and dissolved substances within the oceans. One area off the tip of South Africa has been well studied because numerous eddies are created at the location where the south-moving Agulhas Current meets the subtropical convergence and turns eastward in the Indian Ocean subtropical gyre. In this region, numerous large eddies are formed (**Fig 8-18**), and some of the eddies form warm core rings that spin off into the South Atlantic Ocean, transferring heat and water from the Indian Ocean into the Atlantic Ocean. These eddies travel across the South Atlantic Ocean and to the north and can persist for 3 to 4 years. This transfer of water and heat between oceans is an essential part of the global ocean circulation called the Meridional Overturning Circulation (MOC), which is discussed later in this chapter.

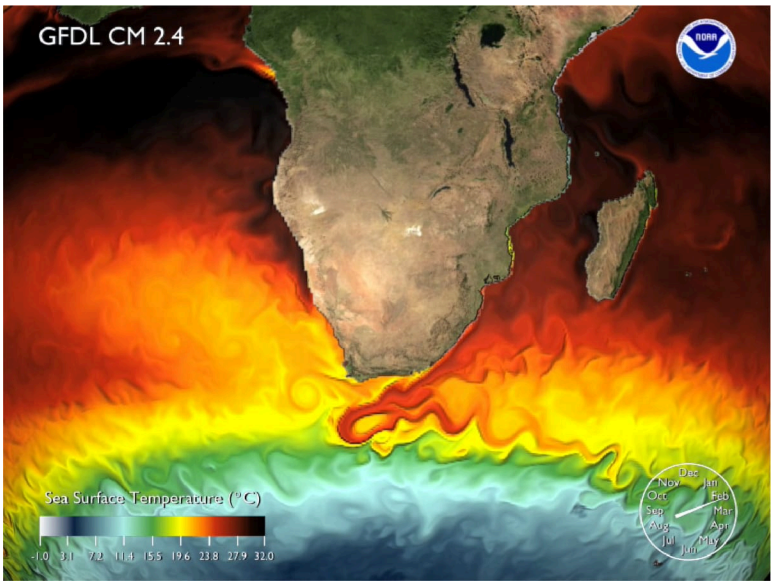

Figure 8-18. Meanders and mesoscale eddies formed at the southern end of the Aghulhas Current. Eddies spin off to the west into the Atlantic Ocean and transport water and heat from the Indian Ocean surface waters to the Atlantic Ocean surface waters. These eddies, similar to Gulf Stream rings, travel to the northwest and persist for several years before being fully mixed with Atlantic Ocean water.

Mesoscale eddies are the subject of considerable ongoing research efforts. However, mesoscale eddies are smaller and slower-moving than atmospheric eddies, and ocean currents are difficult to measure, especially below the surface layer. As a result, the tracking, understanding, and forecasting of ocean “weather,” particularly below the surface layers, will likely always be much more difficult than the observation and forecasting of atmospheric weather.

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8.11: Inertial Currents

Once established by the wind, currents will continue to flow even after the wind stops, until the momentum built up by the wind is dissipated by friction. If there is no other wind or pressure gradient force on the water after the wind stops, the water will flow in a curved path under the influence of the Coriolis effect. If the current is limited to a narrow range of latitude within which the Coriolis effect does not change significantly in magnitude (**CC12**), the water flows in circles. The circular currents are called "inertial currents." Although inertial currents flow even after wind has stopped, they are ultimately wind-driven currents because winds are responsible for starting the water in motion and thus provide the momentum needed to drive the currents.

Inertial currents occur frequently throughout the oceans, but the water rarely moves in perfect circular patterns. In **Figure 8-19**, for example, a circular inertial current is superimposed on a mean drift to the north-northwest. The inertial current and mean wind-driven current were generated by winds blowing at different times, different velocities, and probably different locations. Studies of ocean currents are always complicated by the simultaneous presence of wind-driven and inertial currents, which are caused by different wind events but combine to form the current actually measured at any given place and time. In fact, measured currents are even more complex because geostrophic currents, tidal currents (**Chap. 10**), and wave-induced longshore currents (**Chap. 9**) are also combined with inertial and other wind-driven currents.

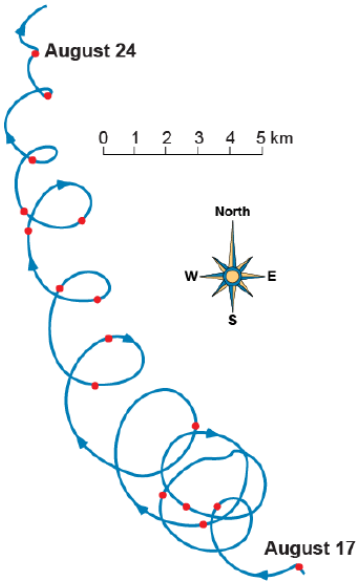

Figure 8-19. Inertial current in the Baltic Sea tracked by a surface float. The inertial motion is superimposed on a mean drift toward the north-northwest. The red dots indicate the float's position every 12 h.

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8.12: Langmuir Circulation

In addition to currents generated by Ekman transport, winds cause a completely different type of near-surface motion in which the upper few meters of water flow in corkscrew-like motions. These motions, called **Langmuir circulation**, are generated when the wind blows across the sea surface at speeds in excess of a few kilometers per hour.

In Langmuir circulation, water travels in helical vortices aligned in the direction of the wind (**Fig. 8-20**). The helical vortices are stacked side by side and rotate in alternating directions like convection cells (**CC3**). Within each cell, water moves across the surface, sinks at a line of downwelling between two cells, moves back under the surface, and rises at a line of upwelling between two cells, while still moving in the wind direction (**Fig. 8-20a**).

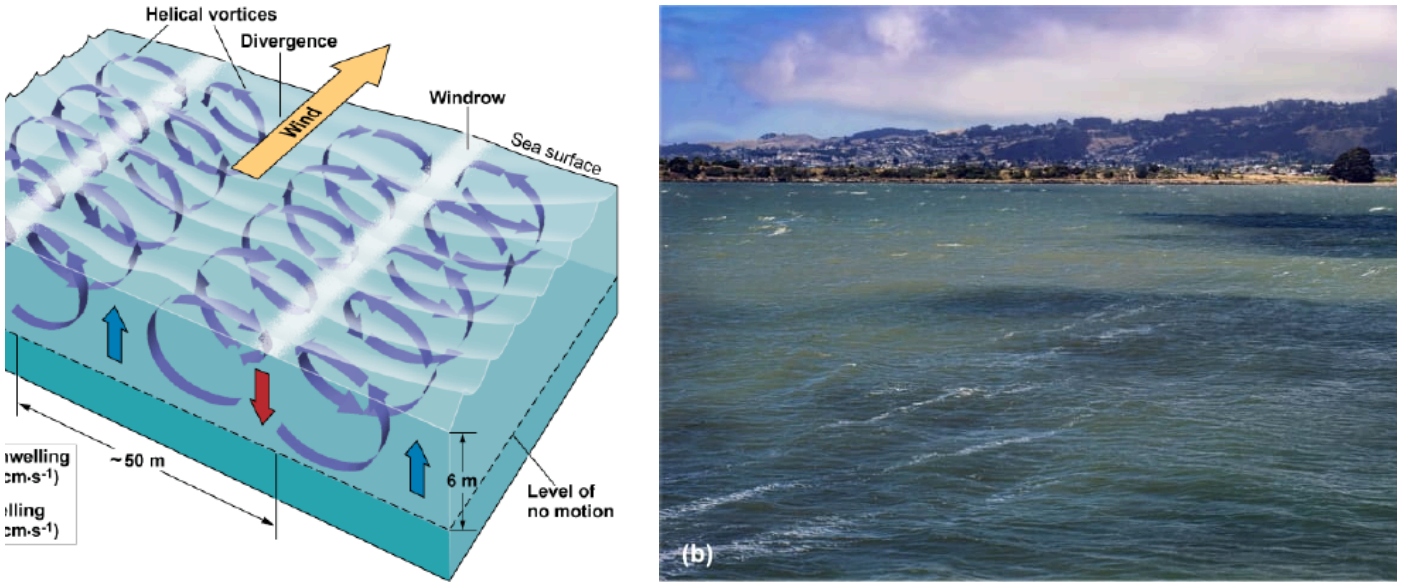

Figure 8-20. Langmuir circulation. (a) Moderate but persistent winds produce corkscrew-like motion of the upper few meters of water. Long linear cells of this helical motion line up alongside each other in the direction of the wind. Adjacent cells rotate in opposite directions, and alternating convergences and divergences form between them. (b) Convergences between Langmuir circulation cells are visible in this photo as the windrows of foam that collect at the convergences when the wind is blowing strongly.

The width and depth of the Langmuir cells are determined by the wind speed and may be limited by shallow water or a shallow pycnocline. Typical cells are a few meters deep and a few meters to about 30 m wide. The top of the downwelling zone between two Langmuir cells is a sea surface convergence. Any oil, debris, or foam that is floating on the surface when Langmuir circulation is established is transported by the circulation to a convergence. The convergence zones are readily observable in lakes or oceans as the linear or nearly linear windrows of floating material that form when strong winds blow (**Fig. 8-20b**). The distance between two of the parallel windrows is equal to twice the width of each Langmuir cell.

Langmuir circulation, together with wave action, is known to be very important in mixing the upper layers of water—a key process in the transport of dissolved gases and heat between the ocean and atmosphere (**Chap. 7**).

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8.13: Thermohaline Circulation

Wind-driven currents dominate water motions in the upper layer of the oceans above the pycnocline. Below the pycnocline, currents are driven primarily by density differences between water masses. Density differences cause water masses to sink or rise to the appropriate density level (**CC1**). In areas where water masses sink or rise, the density distribution with depth, and hence the pressure distribution, is different from that in surrounding areas. Thus, horizontal pressure gradients are formed. Once a water mass has moved vertically to reach its density equilibrium level (**CC1**), it flows horizontally in response to the gradient. Because seawater density is determined primarily by temperature and salinity, these water movements are called "thermohaline circulation."

Thermohaline circulation is difficult to study. Our knowledge about this circulation, especially the details of the complex flow patterns in the deepest water layers and the mechanism and geographical distribution of the upward mixing of water that returns deep water to the surface, is far from complete. Most of our information about these comes from studies of density and other characteristics of the deep-ocean water masses and from models (**CC10**). However, the models are themselves limited by the relatively small amount of data that is available to calibrate and test them.

Depth Distribution of Temperature and Salinity

Ocean waters are arranged in a series of horizontal layers of increasing density from the surface to the ocean floor (**CC1**). Throughout most of the oceans, the layers form three principal depth zones: the surface zone (usually called the **mixed layer**), the pycnocline zone, and the deep zone (Fig. 8-5). The surface zone, which is approximately 100 m thick, is usually of uniform or nearly uniform density. However, a seasonal pycnocline is present in the surface zone in many mid-latitude areas (Fig. 8-21). Water in the surface layer above the pycnocline, or above the seasonal pycnocline if it is present, is continuously mixed or stirred by winds. In the permanent pycnocline zone, water density increases rapidly with depth. This zone extends from the bottom of the surface layer, where water temperature is approximately 10°C, to depths that vary by location between about 500 and 1000 m. In the deep layer below the permanent pycnocline, water density increases slowly with depth.

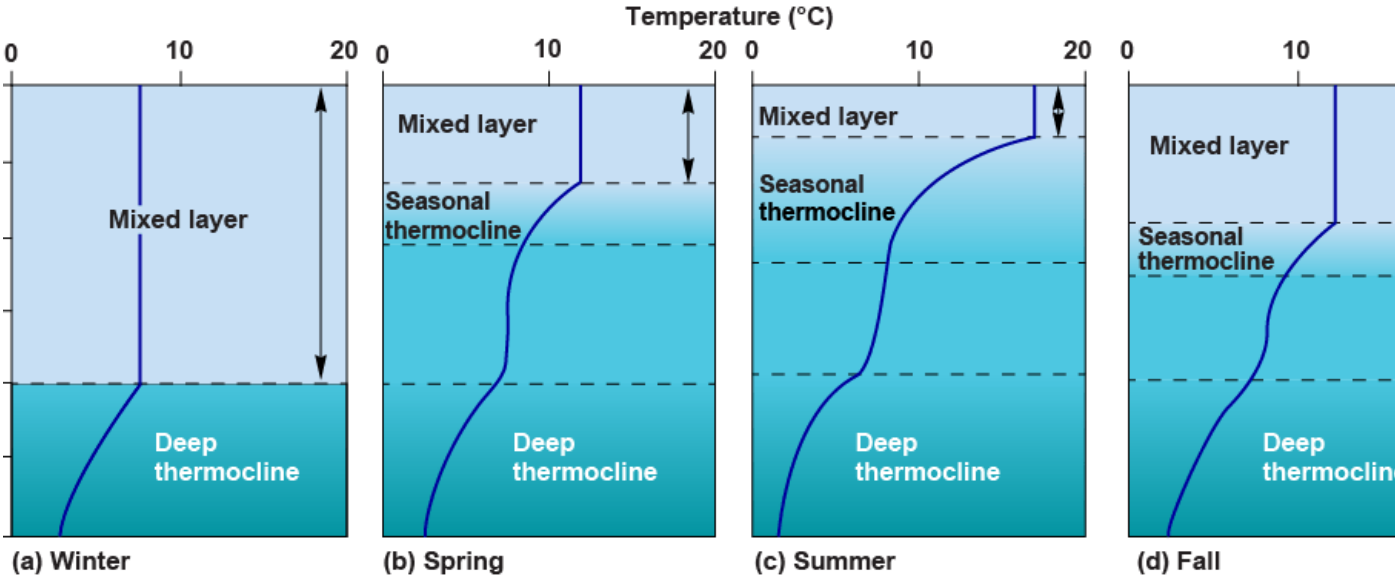

Figure 8-21. In many subtropical and temperate regions, a shallow summer thermocline forms, acting as a barrier to vertical mixing. (a) In winter, the water column is well mixed by the strong winter winds and by convection (due to cooling of surface water) down to the top of the permanent thermocline. (b) In spring, the surface layer is warmed and becomes less dense, forming a shallow thermocline whose depth is determined by wind-mixing depth. (c) As surface water becomes warmer and winds lighter in summer, the seasonal thermocline becomes stronger and shallower. (d) During fall, increased winds increase the depth of the mixed layer, and cooling of the surface water causes convective mixing, which eventually completely destroys the seasonal thermocline.

The marked density increase with depth in the pycnocline zone is due to decreasing temperature, increasing salinity, or a combination of temperature and salinity changes (Fig. 8-5, **CC6**). Where temperature changes cause density to change with

depth, the pycnocline is also a thermocline (**Fig. 8-5b**). Where salinity changes cause density to change with depth, the pycnocline is also a halocline (**Fig. 8-5c**). Haloclines are more important in nearshore waters where freshwater runoff produces surface waters of low salinity.

Figure 8-22 and **Figure 8-23** show the vertical distributions of temperature and salinity, respectively, in the centers of the Atlantic, Pacific, and Indian Oceans. Throughout much of the area between 45°N and 45°S, surface water salinity is actually higher than that of deep waters (**CC6**). The reason is that evaporation exceeds precipitation throughout most of the open oceans in mid- and subtropical latitudes (**Fig. 7-16**). If the water temperature were uniform with depth, the surface waters would be denser than those below and would sink. However, the temperature difference between surface and deep waters is more than enough to offset the density difference due to salinity.

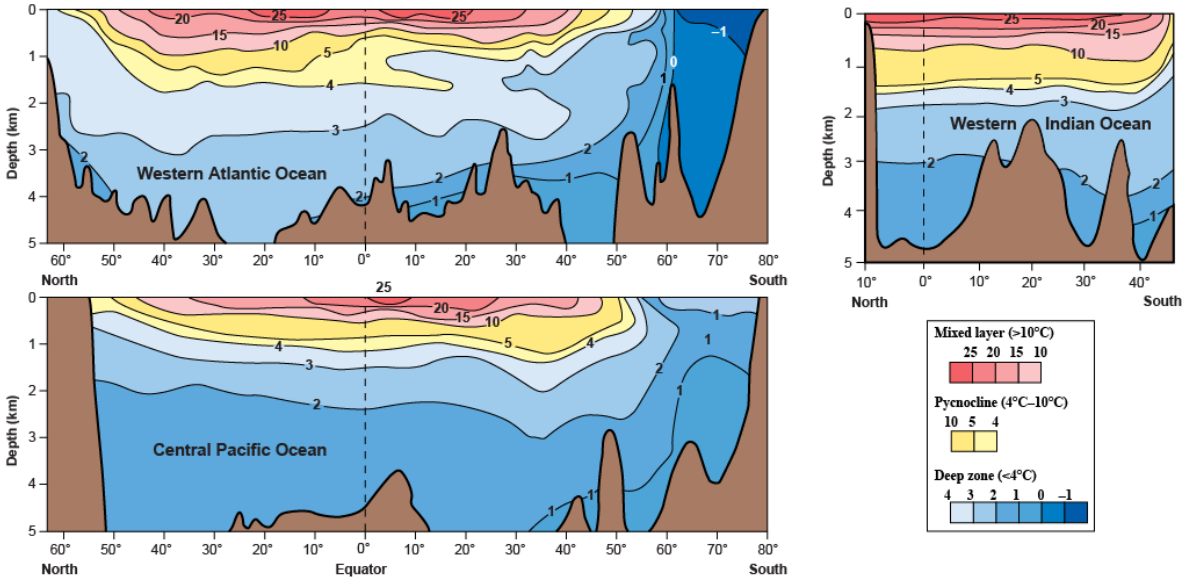

Figure 8-22. North-south profiles showing the distribution of temperature with depth in the open oceans. There are three distinct zones arranged by depth. In the deep zone below the pycnocline, water is cold (below 4°C). Except in high latitudes, there is a warm (>10°C) surface layer that is generally less than 1 km deep. Between the surface and deep zones, the thermocline zone is a region in which temperature decreases rapidly with depth. Around Antarctica, in the North Atlantic Ocean, and to a lesser extent, in the North Pacific Ocean, the warm surface layer and thermocline layer are absent because surface waters in these regions are cooled and sink to form the deep water masses.

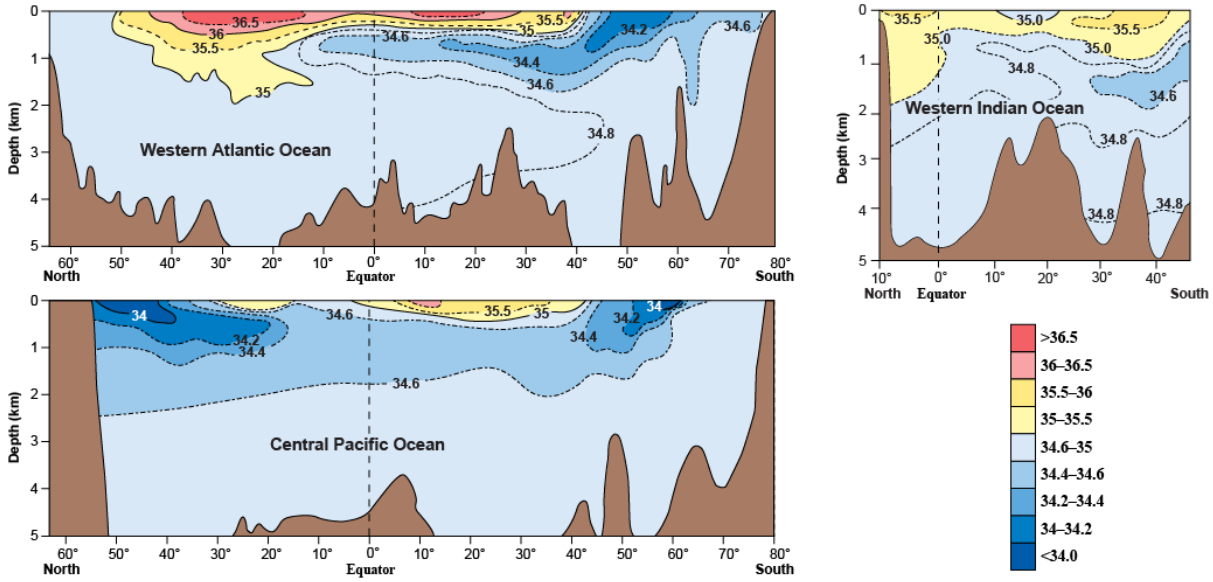

Figure 8-23. North-south profiles showing the distribution of salinity with depth in the deep oceans. This distribution is more complex than the temperature distribution. The surface layer has more variable and generally slightly higher salinity than the deeper water masses, especially in the Atlantic. Salinity is low in the Pacific high latitudes, where precipitation is high, and evaporation is low (dark blue). The influence of the outflow of the warm, high-salinity water mass from the Mediterranean is evident in the tongue of higher-salinity water below 1 km depth between 20°N and 40°N in the Atlantic Ocean.

Within the mixed layer, shallow secondary thermoclines may form during summer in areas where intense solar heating warms surface water, winds are light, and vertical mixing is limited to shallow depths (**Fig. 8-21c**). Seasonal thermoclines are important to biological processes, particularly in coastal waters (**Chap. 13**).

The pycnocline acts as an effective barrier to vertical mixing of water masses. Where density changes rapidly with depth, large amounts of energy are needed to move water either up or down across the density gradient (**CC1**). Consequently, seasonal changes in temperature and salinity caused by changes in solar intensity and rainfall rates (**Chap. 7**) do not generally penetrate below the mixed layer. The pycnocline also acts as a barrier to vertical mixing of dissolved gases and other chemicals. Therefore, deep water is effectively isolated from the mixed layer and atmosphere.

In high latitudes, especially in the northern Atlantic, there is no pycnocline because heat lost from the oceans exceeds heat gained from solar radiation (**Fig. 7-8**). Consequently, surface waters are cooled and their density increases. If cooling is intense enough, the cooled surface water becomes more dense than the water below and sinks. Thus, in high-latitude regions without a pycnocline, deep-ocean water masses are formed as cooled surface waters sink.

The pycnocline layer and deep layer extend throughout each ocean in all but high latitudes. The pycnocline layer is shallower at the equator as a result of upwelling there, and is deeper in some areas, particularly the North Atlantic, and in some areas where complicated tongues of water occur, particularly at depths within and just below the pycnocline zone. The tongues are different water masses because they have their own characteristic temperature and salinity, and they indicate the presence of deep-ocean currents.

Formation of Water Masses

Water density can be increased by decreasing temperature or increasing salinity (**CC6**). Salinity can be increased by evaporation or by ice exclusion (dissolved salts remain in solution in the water left behind as seawater freezes and forms ice crystals). Higher-density water sinks to the depth at which the water below has higher density, and the water above has lower density (**CC1**). It then spreads laterally to form a thin layer extending out from the source area (**Fig. 8-24**).

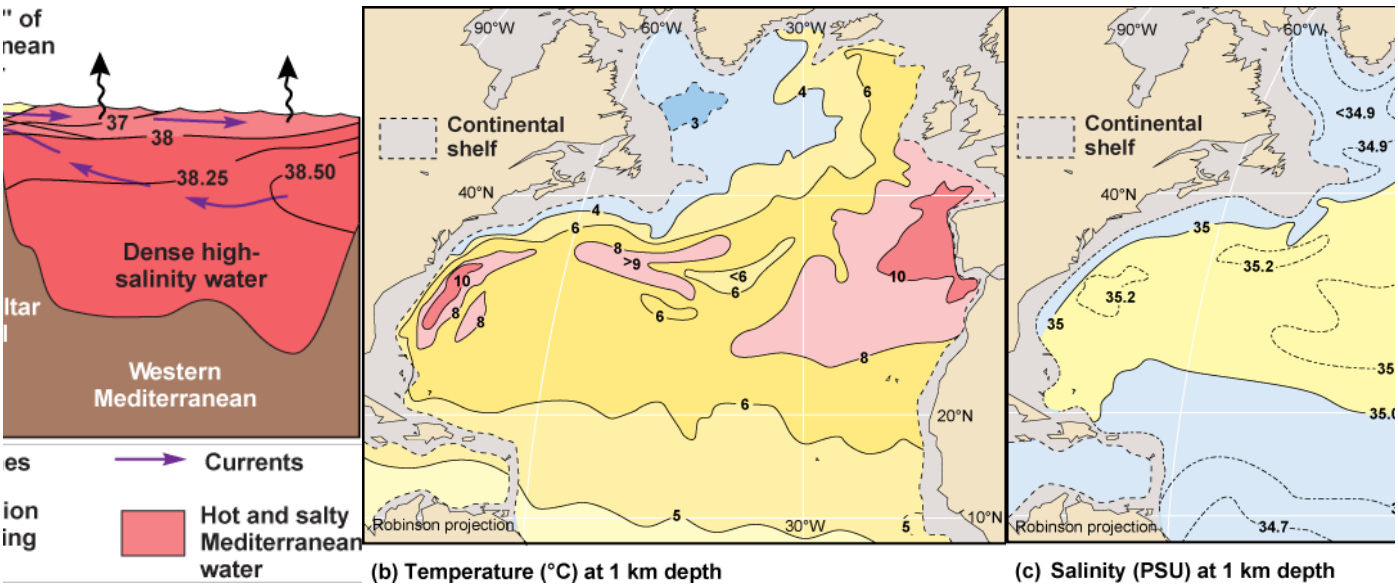

Figure 8-24. (a) The dry, sunny climate in the Mediterranean causes warming and an excess of evaporation over precipitation. The warm, high-salinity water created there spills over the sill at the Straits of Gibraltar into the Atlantic Ocean, where it sinks to its density-equilibrium level at about 1–2 km and spreads out as a distinctive layer. The (b) temperature and (c) salinity characteristics of the Mediterranean water can be detected throughout much of the North Atlantic Ocean.

Water within each such layer is a separate water mass within which salinity, temperature, and therefore density vary only slightly from those of the source water. Water masses can spread horizontally over large areas because the density differences between them inhibit vertical movement and mixing. Less energy is needed for the water mass to flow horizontally than for it to flow or mix vertically because, in a stably **stratified** water column, vertical movements of water must overcome **gravity**. Vertical mixing between water masses of adjacent layers does occur, but the process is generally extremely slow.

Locations of Deep Water-Mass Formation

The densest water masses are created by cooling or freezing of surface waters in only a few locations at high latitudes (**Fig. 8-25**). The densest water is formed by ice exclusion and cooling in the Weddell Sea, a bay on the Antarctic continent opposite the south end of the Atlantic Ocean. This cold, high-salinity water mass, called “Antarctic Bottom Water,” sinks to the deep-ocean floor and is transported eastward around Antarctica. As it sinks, it is partially mixed with other water masses and then moves northward along the ocean bottom into each of the three major ocean basins.

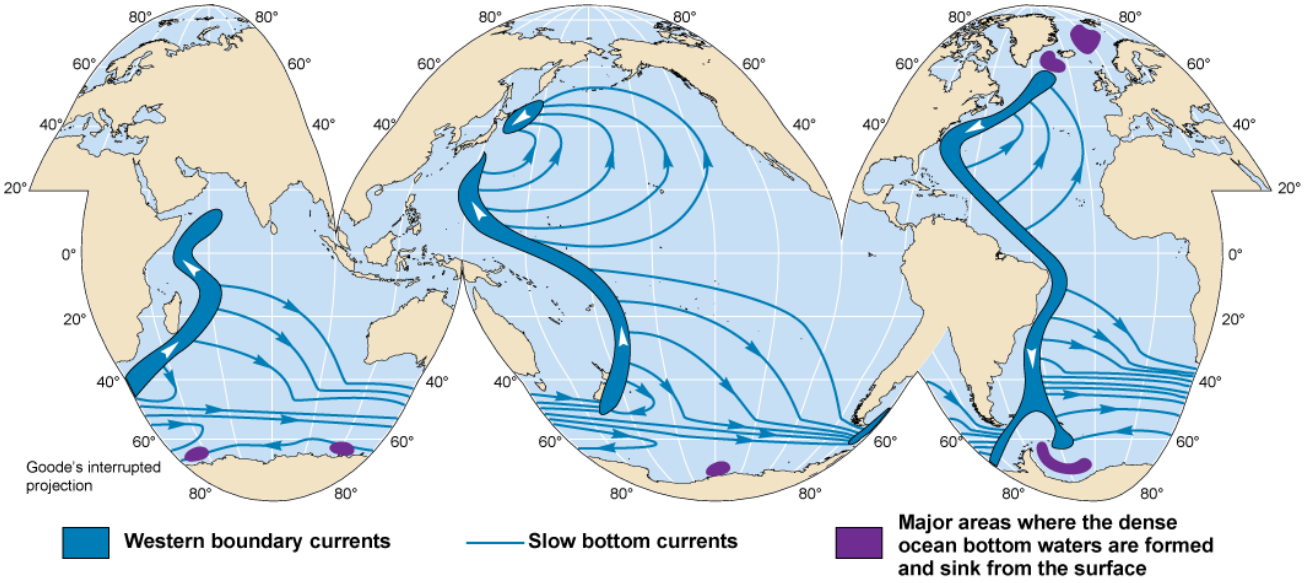

Figure 8-25. The densest water masses that flow along the ocean floor are created by cooling at only a very few locations. Deep water-mass circulation is still not well understood, so this map is only a very general representation of the bottom water flow from each ocean basin through the others. The movement of bottom water within each ocean basin is now known to be far more complex than in this simple depiction. The densest bottom water is formed in the Weddell Sea near Antarctica. It sinks, then flows around Antarctica and northward into each of the oceans. Cold, dense water is also formed near Greenland and in the Norwegian Sea. It flows south until it meets and flows over the more dense Antarctic Bottom Water (Figure 8-29). Bottom currents in each ocean basin are modified by topography, which causes eddy formation. Deep water currents are thought to be weakly intensified on the western boundaries of the oceans (exaggerated in this figure). The western boundary current in the Atlantic Ocean flows southward from the zone of deep water-mass formation near Greenland, and the western boundary currents in the Pacific and Indian oceans flow northward from the Southern Ocean around Antarctica. Return flows of bottom water may exist in some ocean basins but where they exist, they are more complex and diffuse than portrayed here.

No bottom water is formed at the northern end of the Indian or Pacific Oceans. The Indian Ocean does not extend into high latitudes north of the equator. The Pacific Ocean is effectively separated from polar regions by the shallow **sills** between the Aleutian Islands, which mark the southern boundary of the Bering Sea, and by the shallow, narrow Bering Strait that connects the Arctic Ocean and Bering Sea. Cold, dense water formed in the Arctic Ocean sinks and cannot flow into the Pacific Ocean over these shallow areas. In the North Pacific Ocean itself, precipitation rates are high, and low surface salinity prevents the formation of deep water.

The Atlantic Ocean is partially isolated from the Arctic Ocean by shallow ridges between Scotland and Greenland, and the coldest Arctic Ocean deep water cannot readily enter the Atlantic Ocean. However, in the Norwegian and Greenland Sea, intensive cooling forms North Atlantic Deep Water (NADW), which sinks and flows south in vast quantities. It is probably the most voluminous water mass, and its presence can be traced throughout much of the world’s oceans. The Atlantic Ocean has water of higher average salinity than any other ocean for two reasons. First, much of the water evaporated in the tropical Atlantic Ocean is transported across the narrow neck of Central America and deposited as precipitation in the Pacific Ocean. Second, large volumes of high-salinity water are introduced from the Mediterranean Sea (which lies in an atmospheric divergence zone with low rainfall and evaporation exceeding precipitation; **Fig. 8-24**).

The higher salinity and temperature of NADW make it readily distinguishable from Antarctic Bottom Water. Because it is less dense than Antarctic Bottom Water and, therefore, southward-moving NADW flows over the northward-moving Antarctic Bottom Water (**Fig. 8-26**). Although the southward flow of NADW is intensified toward the western boundary, detailed studies show that the southward flow takes place in several broad tongues across most of the width of the Atlantic Ocean, and any return gyral flows in the North Atlantic, such as shown in **Figure 8-25**, are minimal at most.

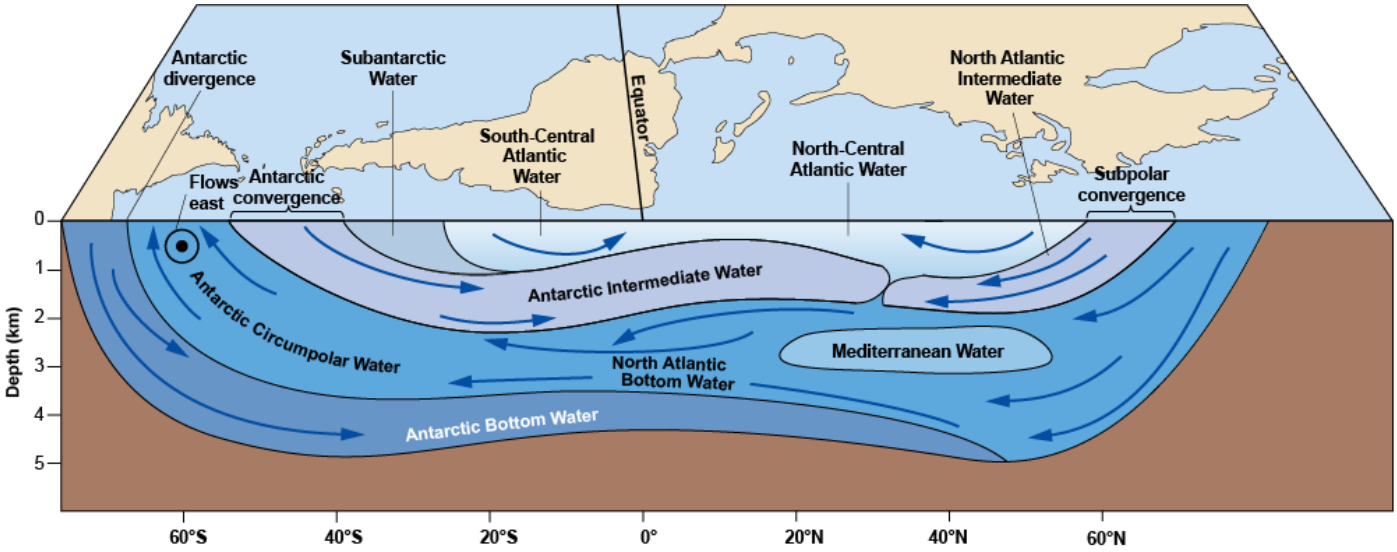

Figure 8-26. A vertical cross-section of the Atlantic Ocean shows the various water masses that form layers at different depths. Antarctic Bottom Water is the densest water mass, and it flows northward from around Antarctica. North Atlantic Bottom Water sinks near Greenland and flows southward over the top of Antarctic Bottom Water until it reaches the surface again in the southern Ocean, where it then mixes with other water masses to form the Antarctic Circumpolar Current. Intermediate-depth water masses are formed and sink at the Antarctic and subpolar convergences. The near-surface layers are more complex. Note the tongue of Mediterranean Water that spreads across the North Atlantic Ocean from the Straits of Gibraltar at about 1 to 3 km depth between 20°N and 55°N.

Most of the Arctic Ocean is covered for much of the year by a floating ice sheet, although the area and thickness of this is rapidly declining (**Fig. 8-27**). Beneath the ice, surface waters have low salinity and are separated from the deep waters by an intense halocline. The low salinity is due partly to river runoff into the Arctic Ocean and partly to ice exclusion. Salt is excluded and transported below the pycnocline during freeze-up in **brines** left by the freezing, but it is not returned in summer when seasonal ice melts. Sea ice contains very little salt, and melting of seasonal ice adds about the same volume of freshwater to the Arctic Ocean surface layer as river runoff does.

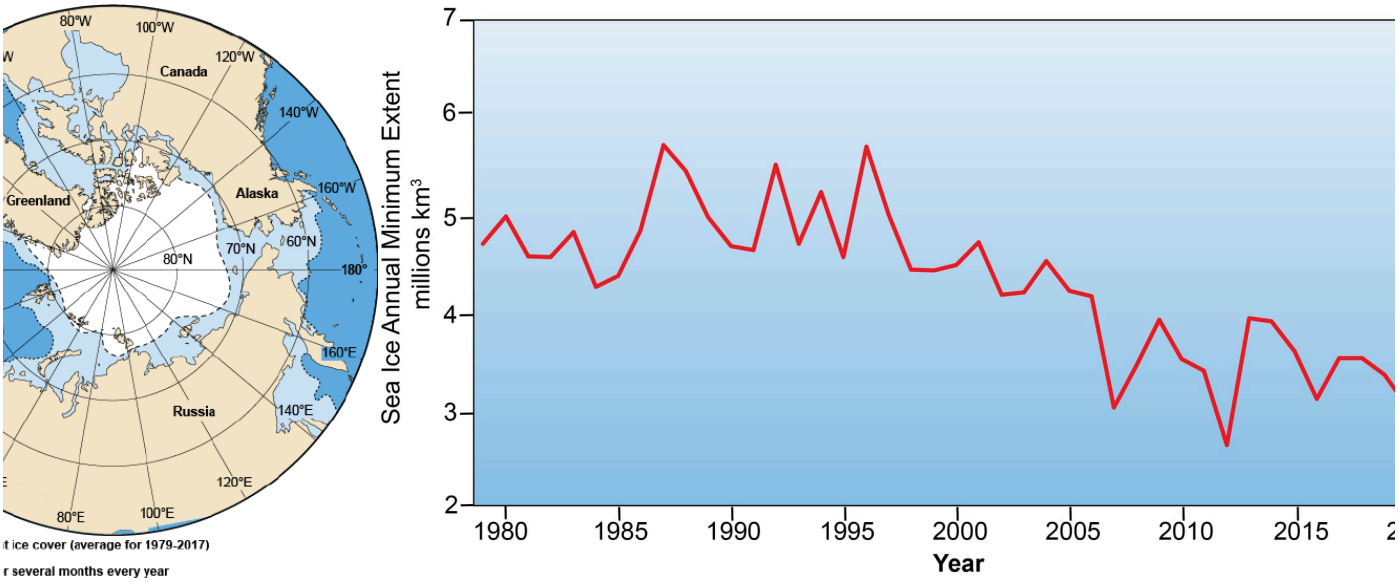

Figure 8-27. (a) Part of the Arctic Ocean is covered by a permanent floating ice pack. Until recent years, only the areas around the edges became ice-free for a few weeks each year. The areas covered by seasonal ice in winter extend far south into the Bering Sea and the Sea of Okhotsk between Russia and Japan, and into the North Atlantic near Greenland and eastern Canada. Most northern European seas are kept ice-free year-round as a result of the warming influence of Gulf Stream water transported into the area. (b) Area of sea ice at its seasonal lowest in September. The area of this permanent ice pack, and of the seasonal ice pack, is becoming smaller from year to year, especially since 2005, and this is believed to be due to the enhanced greenhouse effect. The Arctic Ocean is expected to be ice free in the summer before 2050.

High-salinity brine is also formed by ice exclusion during winter freezing in the continental shelf regions that surround Antarctica. This brine sinks rapidly through the halocline without mixing effectively with the lower-salinity surface water through which it passes. Brine may collect in pools on the continental shelf floor and then drain down the slope. Once in the deep basin below the halocline, the brine eventually mixes with the deep water, most of which enters the Southern Ocean originally from the Atlantic Ocean, and the mixed water mass becomes Antarctic Bottom Water. Most Antarctic Bottom Water is formed in the Weddell Sea section of the Antarctic coast, located south of the southwest corner of the Atlantic Ocean near the tip of South America. Lesser contributions are made at several other locations around Antarctica (**Fig 8-25**).

Deep water-mass movements are affected by the Earth's rotation and the Coriolis effect in the same way that surface currents are. Therefore, deep currents may be intensified along western boundaries of the oceans and may tend to flow in gyre-like motions within the northern and southern ocean basins (**Fig. 8-25**). Eastern boundaries generally have a weak, diffuse flow in the direction opposite to the western boundary flow (**Fig. 8-25**). In addition, deep currents are affected by oceanic ridges and other bottom topography as the flow over and around such topography creates eddy motions.

Water Masses at Intermediate Depths

Intermediate water masses are formed at areas of convergence. For example, sinking of cool, higher-density water at the Antarctic Convergence, and sinking of warm but high-salinity and slightly higher-density water at subtropical convergences, where evaporation exceeds precipitation (**Fig. 8-25**) and form water masses at intermediate depths. Additionally, warm, high-salinity water formed in the Mediterranean Sea, where evaporation exceeds precipitation, is discharged into the Atlantic Ocean, where it sinks to intermediate depths (**Figs. 8-24, 8-26**).

If water continuously sinks to form deep water masses, deep water must be displaced, warmed, and returned to the surface. Cold, deep water is mixed with warmer, less dense water from above by slow vertical mixing between water masses and by more turbulent mixing induced by currents, **internal waves**, and tides. Tides create oscillating currents in the deep waters. Where these currents flow over oceanic ridges or other seafloor topography, eddy formation and turbulent mixing is enhanced, and internal waves can be generated. Vertical mixing is enhanced both at the locations where tidal currents meet rough seafloor topography and at locations where internal waves break as the seafloor shallows on the continental slope or on oceanic ridges and other seamounts. Thus, vertical mixing of deep water upwards back to the surface layers takes place very slowly by **diffusion**, except where seafloor topography interacts with deep currents and internal waves. The locations where such interaction with topography occur are numerous and located primarily at oceanic ridges, seamounts, and on continental slopes. At these locations is known to be vertically mixed with shallower water masses by turbulence at rates far above that in the open ocean, and this also can occur in pulses. This is most likely the dominant mechanism for returning bottom waters back to the surface layers. Water from intermediate depths may also return to the surface by upwelling at divergences such as the Antarctic Divergence (**Fig. 8-26**).

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8.14: Ocean Circulation and Climate

Circulation of water in the oceans can be likened to a giant conveyor belt. Water is cooled near the poles; transported through the deep oceans, where it mixes with other water; returned to the surface far from where it sank; and warmed and transported back to a location where it cools and enters the cycle again.

The ocean circulation system carries heat from one part of the planet to another and transfers dissolved chemicals between surface and deep-water layers. Consequently, changes in circulation can cause and be caused by changes in climate, and they can cause changes in the distribution of dissolved chemicals that directly affect the biological **communities** of the oceans. Ocean circulation also carries some of the excess carbon dioxide from **fossil fuel** burning into the deep oceans (**CC9**). Therefore, deep-ocean circulation research is of great interest, particularly for global climate change studies.

Meridional Overturning Circulation: The Conveyor Belt

The Meridional Overturning Circulation (MOC), which is often referred to as the "conveyor belt circulation," is a closed-loop circulation of water that travels through all of the world's oceans. The general features of the MOC are depicted in **Figure 8-28**. The complete circuit takes an average of about 1000 years, and the amount of water transported is enormous—about 20 times the combined flow of all the world's rivers. The MOC is usually considered to start in the North Atlantic near Greenland and Iceland, where surface water is cooled by the cold air mass that flows from the Canadian Arctic. The cold, high-density water sinks to form North Atlantic Deep Water (NADW), which then flows south through the deep Atlantic Ocean, around Africa, where it flows over and mixes with the deep water formed near Antarctica. The deep water is transported around Antarctica, then north into the Indian and Pacific Oceans. The water mixes progressively toward the surface. The details of where and how this mixing takes place are not yet well known, but it is thought to take place slowly throughout most of the oceans and more quickly through turbulent mixing in some areas where currents and internal waves interact with seafloor topography. Eventually, the deep water mixes with surface waters and enters the mixed-layer current system. Complex exchanges and movements of surface water eventually return surface water to the Atlantic Ocean, replacing the water that originally formed NADW. This system is instrumental in transporting heat from low latitudes to high latitudes of the North Atlantic Ocean.

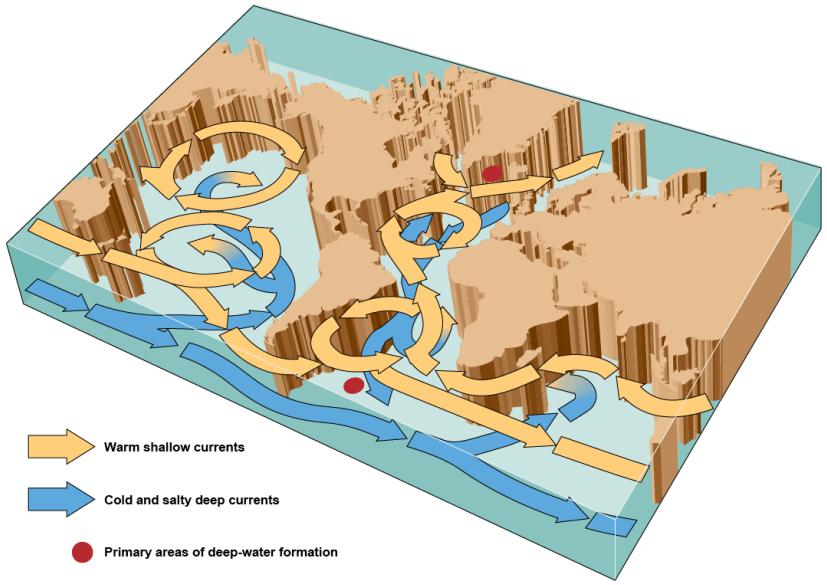

Figure 8-28. A simplified depiction of the Meridional Overturning Circulation (MOC), or "conveyor belt," that starts with the formation of North Atlantic Deep Water in several locations near Greenland.

Surface water in the North Atlantic Ocean is warmer than NADW. As the Gulf Stream flows into the northeastern Atlantic Ocean near Europe, prevailing westerly winds blow over it, transferring heat and moisture into the atmosphere. The westerly wind air mass releases its heat and moisture over Europe, causing Europe's climate to be extremely mild and wet in comparison with the climates of other land areas at the same latitude (**Chap. 7**). Thus, the MOC transfers heat from the Indian Ocean and Pacific Ocean to the North Atlantic region near Europe. At the same time, the MOC transports dissolved nutrients from the Atlantic Ocean

to surface waters of the Indian and Pacific Oceans.

The MOC Climate Switch

Ocean **sediment** records have shown that, during the last **ice age**, which peaked about 18,000 years ago, the MOC generally operated more slowly and weakly than it does today, and varied in strength. The variances may have occurred abruptly at times. Some of those abrupt changes may have almost entirely turned the MOC off or back on again during the past tens of thousands of years. It is believed that such abrupt changes have, at least on some occasions, coincided with very abrupt (on the scale of decades) changes in climate. Evidence suggests that variations in the MOC circulation, at least during the cold, ice-age period, may have been **chaotic** (**CC11**) and that ocean circulation may have switched periodically between a very weak MOC and a strong MOC, which is seen today.

About 15,000 years ago, the last ice age ended, although it may be that the current warm period is just a warm interglacial interval similar to other interglacial periods that have interspersed the ice age in the past. Nevertheless, about 15,000 years ago, the Earth's average temperature increased by about 6°C in as little as 100 years. This abrupt change was accompanied by a 20% increase in atmospheric carbon dioxide concentration and a substantial, but not yet well quantified, increase in the intensity of the MOC.

During the past 13,500 years, the Earth's climate has been warmer than at any other time during the past million or more years. However, this warm period was interrupted by a short cold period, the Younger Dryas period or "Little Ice Age", that occurred between approximately 12,900 years ago and 11,500 years ago. At the beginning of the Younger Dryas period, western Europe's climate cooled dramatically by about 5°C within a matter of decades or less and then returned just as abruptly to its former warm condition at the end of this period. Apparently, the Younger Dryas cold climate occurred when the MOC was severely slowed or stopped. The reason for the temporary slowing or stoppage of the MOC is not known, but might be related to the flow of meltwater from **glaciers**.

During the early postglacial period, about 13,000 years ago, glaciers extended far to the south, and most meltwater from the North American ice sheet probably drained down the Mississippi Valley to the Gulf of Mexico. By 13,000 years ago, as the glaciers continued to melt, a lake called Lake Agassiz covered much of Manitoba, western Ontario, northern Minnesota, eastern North Dakota, and Saskatchewan. At its greatest extent, Lake Agassiz may have covered as much as 440,000 km2, larger than any currently existing lake in the world. It is believed that the cold, fresh meltwater that was collected in this lake emptied rapidly either through the McKenzie River to the Arctic Ocean or through the Saint Lawrence River to the North Atlantic, or perhaps both. Because it was fresh or very low salinity water, it would have floated on the ocean water to form a stable, low-density surface layer over a large area of the Arctic Ocean and North Atlantic Ocean. The low-salinity surface layer would have acted as a virtual cap, severely restricting the formation of NADW and slowing the MOC. Within several hundred years after glacial meltwater first flooded out of Lake Agassiz in large volumes the freshwater flow to the North Atlantic probably diminished because, by that time, most glaciers had finished melting. At that point, NADW again formed, the MOC resumed, and Europe's climate warmed rapidly.

We are not sure whether a number of additional abrupt climate changes that have occurred during the past 10,000 to 12,000 years coincided with significant changes in the rate of water transport within the MOC, but it appears that at least some such abrupt climate changes did coincide with such changes. We cannot yet determine for certain whether changes in the MOC caused these additional abrupt climate changes, or vice versa.

Recent studies of the MOC suggest that the rate of formation of North Atlantic Deep Water and thus the rate of flow in the MOC exhibits strong short-term variability that makes it very difficult to assess whether there are any long-term trends in its strength. However, some estimates suggest that the MOC may have slowed down, perhaps as much as 30% during the past several decades. This slowdown, if it can be confirmed, may be related to climate change. If so, and if this indicates that the MOC will continue to slow down as the planet warms due to anthropogenic climate change, one possible result could be the abrupt (years to decades) onset of a cold climate period in Europe, similar to the Younger Dryas period, and to other abrupt climate changes elsewhere. However, modeling studies (**CC10**) suggest that climate change is likely to slow the MOC but not to cause it to slow so much that catastrophic changes, including drastic cooling of Europe, will occur within the next century or two.

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8.15: Tracing Water Masses

Although direct measurements have been made of deep-ocean currents, most of what we know about deep-ocean circulation was learned from studies of the horizontal and vertical distribution of water masses. Water masses are most often characterized by their temperature and salinity, but the concentrations of several dissolved constituents are also used to distinguish and trace water masses (**CC6**).

Conservative and Nonconservative Properties

Conservative properties of seawater are properties that are not changed by biological, chemical, or physical processes within the ocean. A conservative property can be changed only by mixing with other water masses that have different values of that property, or by processes that occur at the ocean surface, or where rivers or other sources, such as **hydrothermal vents**, enter the oceans. For example, salinity, a conservative property, can be changed at the ocean surface by evaporation or precipitation, or at river mouths by the introduction of freshwater. Small salinity changes can also occur at hydrothermal vents, where high-salinity waters are emitted. Similarly, temperature is a conservative property until water contacts the atmosphere or seafloor, with which it can exchange heat.

Dissolved constituents of seawater, such as Na^+ and Cl^-, whose concentrations are not significantly affected by biological or chemical uptake or removal from ocean water, are also conservative properties. In contrast, concentrations of constituents such as oxygen, carbon dioxide, phosphate, and many trace metals are substantially altered by biological and chemical processes that remove and release them to solution within the ocean. These are nonconservative properties.

Conservative properties are particularly useful in tracing water masses because they can be used to identify the masses as they are formed, to trace their transport through the oceans, and to determine how they mix with other water masses. When two water masses mix, the value of the conservative property in the mixed water mass is determined by the proportions in which the two water masses have mixed. For example, consider the mixing of two water masses, one with a salinity of 35 and the other with a salinity of 37. If the salinity of the mixed water mass is 36, the mixture must consist of equal volumes of the two water masses (2 volumes of 36 = 1 volume of 37 + 1 volume of 35). If the salinity of the mixture is 35.5, the mixture must consist of three parts water of salinity 35 and 1 part water of salinity 37 (4 volumes of 35.5 = 3 volumes of 35 + 1 volume of 37).

TS Diagrams

We can identify different water masses by plotting temperature against salinity in what are called “TS diagrams” (**Fig. 8-29**). Water samples from the same water mass have the same salinity and temperature and therefore will appear as a single point on a TS diagram. Samples from water masses that have different temperature and salinity appear as separated points. When two water masses are mixed, the temperature and salinity of the mixed water lie along a straight line drawn between the two points representing the original water masses. The location of the TS value for a mixed water sample on the line between the TS values of the two water sources indicates the relative proportions of the two original water masses in the mixture.

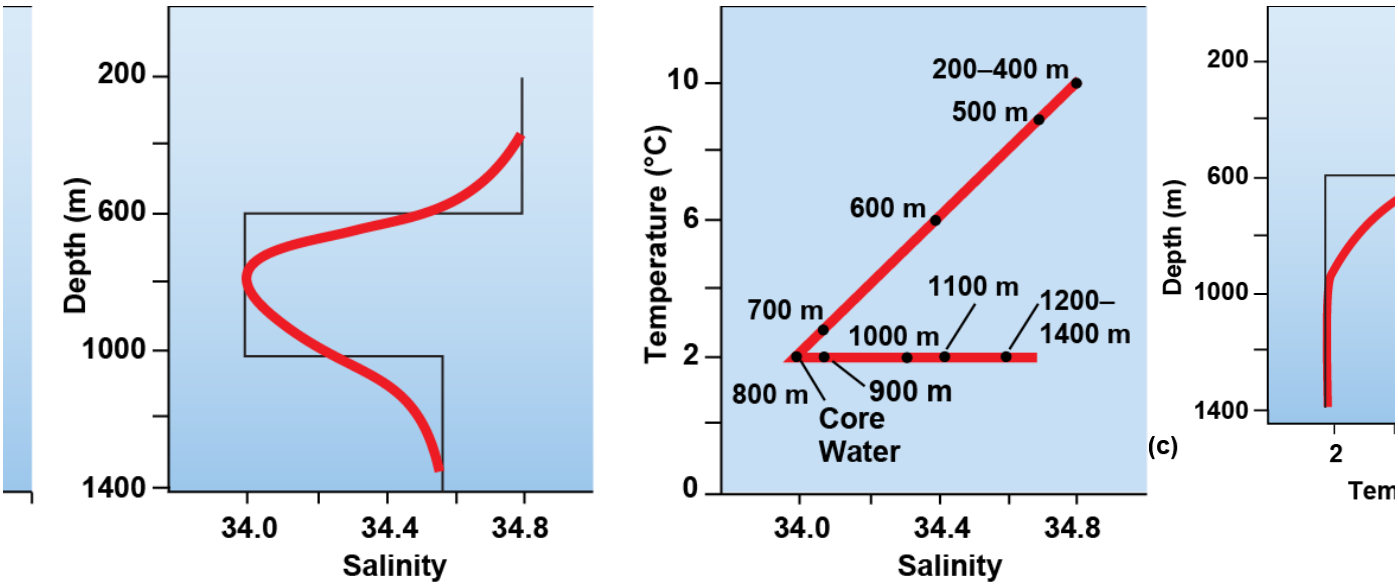

Figure 8-29. TS diagrams are used to trace water masses. In this figure, hypothetical vertical temperature and salinity profiles are used to construct examples of TS diagrams. (a) A simple three-layer system. All TS points plot at one of three locations, representing the three water masses. (b) If the three layers mix, but some of the unmixed mid-depth water mass is still present (at 700 m depth), the TS points plot on two straight lines that connect the three original water-mass characteristic points. (c) Even if all of the mid-depth water mass is mixed with at least some of the other two water masses, a TS plot can be used to identify the temperature and salinity characteristics of the original unmixed water mass. (d) A typical TS diagram for a station in the South Atlantic Ocean.

When three water masses mix in such a way that no unmixed sample of one of the three water masses remains, some of the mixed water samples will have TS values that are no longer on straight lines between the three original TS points (**Fig. 8-29c**). However, in such cases, the temperature and salinity of the original water mass, which now exists only as a part of mixtures, can often be deduced by extrapolation of the straight-line portions of the TS curve (**Fig. 8-29c**).

A TS diagram for water samples taken at different depths at a single station in the South Atlantic Ocean is shown in **Figure 8-29d**. At this station, the principal water masses are warm, saline surface water; North Atlantic Central Water; water that is a mixture of surface water and North Atlantic Deep Water; Antarctic Intermediate Water; North Atlantic Deep Water; and Antarctic Bottom Water. We need not be concerned with how these water masses acquired their different temperature and salinity characteristics. What is important is that, once the source properties are known, the presence of a particular water mass in the area for which the TS diagram is drawn can be determined. Note that we can detect the presence of Antarctic Intermediate Water in the diagram, even though no unmixed Antarctic Intermediate Water is present at the station. We can also calculate from the diagram the percentage of Antarctic Intermediate Water present in the mixed water at any depth. For example, at 800 m, where the proportion of this water mass is greatest, Antarctic Intermediate Water is about 55% of the mixture.

By drawing TS diagrams for many different stations across the oceans, oceanographers can trace movements of individual water masses and investigate mixing processes at different points within the oceans.

Tracers

Salinity and temperature are excellent tracers, but they tell us only which water masses are mixed and where the mixed water is transported. Without other information, these tracers cannot tell us how fast the water masses move and mix. Several different dissolved components of seawater are now used in conjunction with temperature and salinity to trace ocean water masses and to provide information on their rate of movement.

In some applications, nonconservative properties, such as oxygen and carbon dioxide concentrations, can be used as tracers. When water sinks below the pycnocline and is no longer in contact with the atmosphere, the concentrations of these gases are altered by the decomposition of organic particles, which consumes oxygen and releases additional carbon dioxide to solution. Therefore, if we know the oxygen and carbon dioxide concentrations in water masses when they sink below the mixed layer, changes in those concentrations can provide information about the relative ages (since the time when they sank) of water masses in different parts of the oceans.

Human activities have provided a number of useful tracers during the past half century. These tracers include **radionuclides** created by nuclear weapons testing, and certain synthetic organic compounds, particularly **chlorofluorocarbons (CFCs)**. Some radionuclides that do not occur naturally are now found dissolved in ocean water in extremely small concentrations. They are derived primarily from nuclear bomb tests in the atmosphere, which peaked in the 1950s. These radionuclides are good tracers because the time and location of their introduction to the oceans is well known, and they decay at known, constant rates (**CC7**).

Tritium (3H), a **radioactive isotope** of hydrogen, is a particularly useful tracer. Tritium is produced naturally by cosmic rays in the atmosphere. Thus, it occurs naturally in ocean water, but only at extremely small concentrations. In water unaffected by human releases, approximately one of every 1019 atoms of hydrogen is a tritium atom. Releases of tritium during the nuclear bomb testing era of the mid-twentieth century raised the tritium concentration in surface seawater to a concentration of approximately 6 atoms of tritium in 1018 atoms of hydrogen. Tritium reacts quickly with oxygen in the atmosphere to form tritiated water, which enters the oceans in rainfall. Because the tritium is incorporated in the water molecule itself, it is a perfect tracer for water masses. Tritium has a **half-life** of only approximately 12 years, so bomb-produced tritium will continue to be useful as a tracer for only a few decades, until all the tritium introduced by nuclear bomb testing has decayed.

In the Northern Hemisphere, most tritium was produced by nuclear tests in the Soviet Arctic. Therefore, tritium concentrations are particularly high in water flowing into the Atlantic Ocean from the Arctic Ocean. This tritium is now distributed throughout the mixed layer of the oceans and is currently being transported into the deep oceans with sinking cold water masses created in polar regions. The progress of deep water-mass formation has been monitored using tritium as a tracer (**Fig. 8-30**).

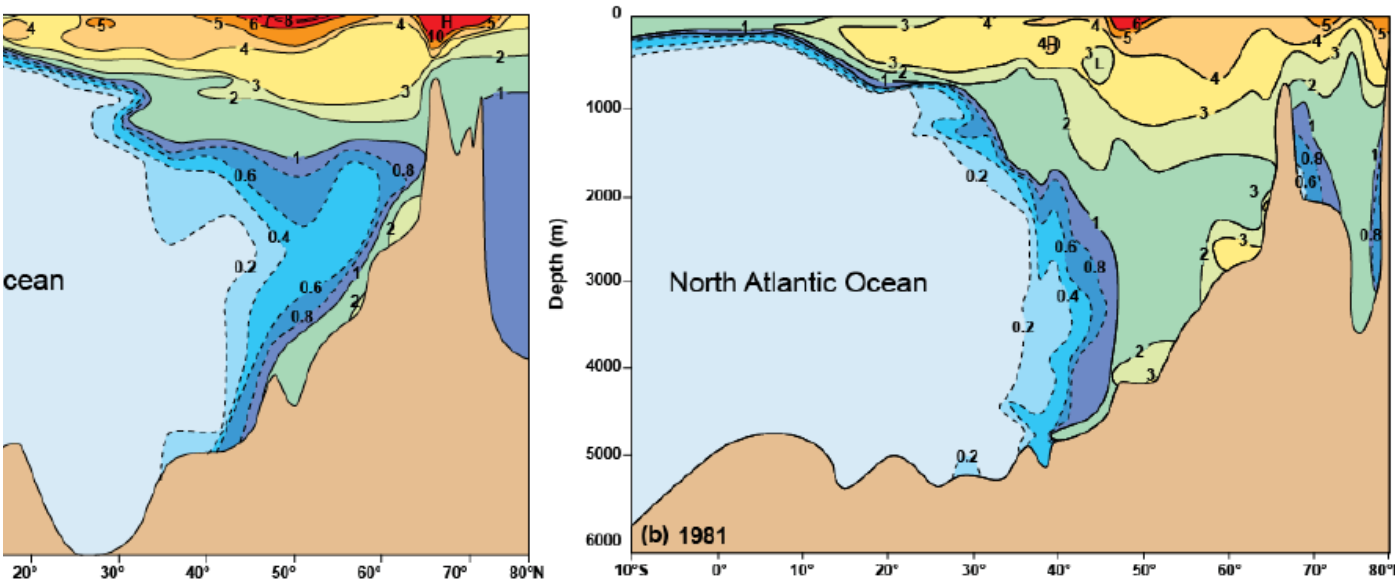

Figure 8-30. (a) Tritium distribution with depth in the western North Atlantic Ocean (1972–1973). Released primarily by nuclear bomb tests in the 1950s, tritium has spread throughout the ocean surface layers and is being transported steadily into the deep oceans with North Atlantic Deep Water, which is formed by the cooling of surface waters in the North Atlantic. (b) Distribution in 1981. The values are corrected for the radioactive decay that occurred during the 9 years between the two surveys. The strong source in the surface layer at 50°N is the outflow from the Arctic Ocean where the Soviet Union did most of its nuclear bomb testing. Comparing the two diagrams shows the progressive movement of tritium into the deep layer as North Atlantic Deep Water continuously forms, sinks, and moves southward.

Chlorofluorocarbons (CFCs), which were used as refrigerants and are thought to be responsible for **ozone** depletion in the atmosphere (**Chap. 7**), are also excellent water-mass tracers (**Fig. 8-31**). These compounds do not occur naturally. They are dissolved at very low concentrations in seawater and are not readily decomposed by biological processes.

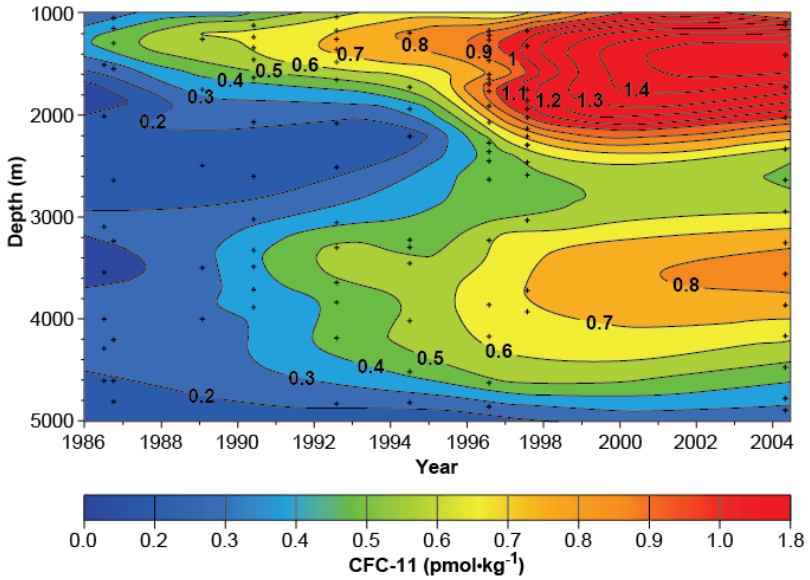

Figure 8-31. The concentrations of CFC-11, a specific chlorofluorocarbon compound, have been measured throughout the water column to the east of Abaco Island in the Bahamas, periodically between 1986 and 2004. This figure plots the vertical distribution of the concentrations of CFC-11 in this area as a function of time. The data show the arrival at this subtropical location of CFC-11 in recently formed deep water masses originating from the high latitude, deep water mass formation areas in the polar and subpolar regions of the North Atlantic (**Figs. 8-26, 8-28**). There are three distinct water layers seen arriving in this area, centered at about 1200 m, 1800 m, and 4000 m. The deepest layer is deep water formed from the outflow of the Arctic Ocean, and the two shallower layers are both deep water formed near Greenland. The 1800 m layer arrived near Abaco later than the shallower layer, and appears to reflect a small reduction in the temperature of deep water formed near Greenland that had been observed since about a decade earlier.

Unfortunately, for most water-mass tracers, **in situ** instrumentation is not sensitive enough, and thus, water samples must be collected for laboratory analysis. Because the cost of ship time required to collect water samples from deep within the water column is very high, sampling cannot be as intensive either spatially or temporally as would be desirable in most instances. Thus, many studies rely on mathematical modeling (**CC10**) using the limited data that are available.

Studying the recent and ongoing processes of deep water mass formation is extremely important to our understanding of the effects of fossil fuel burning on global climate. Estimates suggest that approximately 30% of the carbon dioxide released by burning fossil fuels since the Industrial Revolution in the mid-1800s is now dissolved in ocean water. However, there is still considerable uncertainty about the accuracy of the estimates. Ocean water that sinks below the mixed layer carries carbon dioxide from the atmosphere into the deep oceans, where it is locked away from contact with the atmosphere for hundreds or thousands of years (or longer if it is incorporated in carbonate sediments; **Chap. 6**). Tracer studies will help us to determine how much carbon dioxide from fossil fuels has already been transported safely, at least for now, below the pycnocline and to estimate how quickly additional carbon dioxide will follow it.

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8.16: Chapter Summary

Energy Sources.

Winds are the energy source for most currents in the ocean's upper few hundred meters. Deeper currents and most vertical movements of ocean water are caused by thermohaline circulation, which is driven by differences in the density of water masses due to temperature and salinity differences.

Wind-Driven Currents.

Winds blowing across the sea surface transfer kinetic energy to the water through friction and cause wind-driven currents with speeds generally about 1% to 3% of the wind speed. Currents continue after winds abate until their momentum is eventually lost because of internal friction. Currents also continue because winds transport water, cause the sea surface to slope, and create horizontal pressure gradients that drive water motion. Current speed and direction are modified by friction, the Coriolis effect, horizontal pressure gradients, and land and seafloor topography.

Ekman Motion.

Surface water set in motion by winds is deflected *cum sole* relative to the wind direction by the Coriolis effect. Wind energy is transferred down into the water column as each water layer transfers its energy to the layer below. If winds blow consistently for long enough over deep water, an Ekman spiral develops in which the surface current is deflected 45° *cum sole*, current speed decreases with depth, and current direction is progressively deflected *cum sole* with depth. Mean water transport (Ekman transport) in the layer within which the Ekman spiral develops is at 90° *cum sole* to the wind.

Wind-driven currents rarely reach deeper than 100 to 200 m. Where the seafloor is shallower than the Ekman spiral depth, friction reduces the surface water deflection below 45° and the mean transport below 90°. If a pycnocline is present within the Ekman spiral depth, downward transfer of wind energy is inhibited, and wind-driven currents do not occur below the pycnocline.

Geostrophic Currents.

Ekman transport causes surface water to "pile up" in some areas. The resulting sloping sea surface results in a horizontal pressure gradient within the upper several hundred meters. Water flows on such gradients and is deflected by the Coriolis effect. The deflection continues until the flow is along the gradient (parallel to the pressure contours) at a speed where the Coriolis deflection is just matched by the pressure gradient force. This is geostrophic flow.

Ocean Surface Currents.

Trade winds drive surface waters generally west in the Equatorial Currents. Westerly winds drive surface waters generally east in the subtropics. Continents deflect the flow north or south. Thus, current gyres are created between trade wind and westerly wind latitudes in each hemisphere and ocean. These gyres flow clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. Ekman transport by trade winds and westerlies piles up water in the gyre center, creating horizontal pressure gradients that maintain geostrophic gyre currents even when winds cease. Western boundary currents are faster, deeper, and narrower than eastern boundary currents.

In the high latitudes of the North Atlantic and Pacific Oceans, counterclockwise subpolar gyres are complex, weak, and variable. At high southern latitudes, no continents block the Antarctic Circumpolar Current that flows westward around Antarctica.

Upwelling and Downwelling.

Upwelling occurs where Ekman transport moves surface waters away from a divergence or coast. Downwelling occurs where Ekman transport adds to the surface layer depth at convergences or coasts. Upwelling areas are usually highly productive because

upwelling can bring cold, nutrient-rich water from below into the mixed layer, where nutrients are essential for phytoplankton growth. Divergences and upwelling occur in part of the area between the North and South Pacific Equatorial Currents, around Antarctica, and in many coastal locations. Convergences occur in the centers of ocean gyres and in many coastal regions. Coastal upwelling is more prevalent inshore of eastern boundary currents than inshore of western boundary currents.

Coastal Currents.

Local winds determine coastal current directions and they may be opposite those of the adjacent gyre currents. Coastal currents are variable and are affected by freshwater inflow, friction with the seafloor, steering by seafloor topography and coastline, and tidal currents.

Eddies.

Ocean eddies are similar to, but smaller than, the atmospheric eddies seen on satellite images as swirls of clouds. Satellites can be used to observe some ocean eddies. Gulf Stream rings are mesoscale eddies 100 to 300 km wide that form cold-core (counterclockwise-rotating) or warm-core (clockwise-rotating) rings when a meander of the Gulf Stream breaks off. The rings, which may last several months, may reattach to the Gulf Stream as they drift slowly south. Mesoscale eddies are present throughout the oceans, generally have slower currents than Gulf Stream rings, may extend to the deep-sea floor, and are 25 to 200 km wide.

Inertial Currents.

Once established, wind-driven currents that continue to flow and are deflected into circular paths by the Coriolis effect are called “inertial currents.”

Langmuir Circulation.

Strong winds can set up a corkscrew-like Langmuir circulation aligned in the wind direction. Langmuir cells are a few meters deep and about 30 m wide, and they lie side by side. Foam or floating debris collects in the linear downwelling regions between cells.

Thermohaline Circulation.

When surface water density is increased by cooling or evaporation, the water sinks and spreads horizontally at a depth where the water above has a higher density and the water below has a lower density. Throughout most of the oceans, the water column consists of a uniform-density mixed surface layer (about 100 m deep), a permanent pycnocline zone (extending from the bottom of the mixed layer to a depth of 500 to 1000 m) in which density increases progressively with depth, and a deep zone in which density increases slowly with depth. Most high-latitude regions where surface waters are cooled lack a pycnocline. Secondary, temporary pycnoclines can develop in the mixed layer in summer. Pycnoclines act as a barrier to vertical mixing.

Deep water masses are formed by cooling and ice exclusion at high southern latitudes, in the Arctic Ocean, and at high latitudes in the North Atlantic Ocean. Antarctic Bottom Water is formed primarily in the Weddell Sea and flows north as the deepest and densest water layer. North Atlantic Deep Water is formed in the Greenland and Norwegian Seas and flows south. Water masses at intermediate depths are formed by the sinking of cooled water at the Antarctic Convergence and of warm, high-salinity water produced by evaporation at subtropical convergences and in the Mediterranean Sea. Deep and intermediate-depth water masses are subject to the Coriolis effect, and thus, currents in the deep layer tend to be intensified along western boundaries and tend to flow in gyres. They are also affected by seafloor topography, which generates eddies that enhance vertical mixing.

Ocean Circulation and Climate.

Ocean circulation transfers heat from the tropics toward the poles, moderating mid- and high-latitude climates. North Atlantic Deep Water forms near Greenland and Iceland. It sinks and spreads south in the deep Atlantic and then north into the Pacific and Indian Oceans. It mixes progressively upward to the surface layer, which is warmed and eventually transferred back to the North Atlantic by the Gulf Stream.

This Meridional Overturning Circulation (MOC) has varied in intensity and switched off and on in the past. Periods when it has not operated seem to be associated with colder climates, especially in Europe. The change to a colder climate appears to take place abruptly when the MOC is switched off.

Tracing Water Masses.

Water masses are characterized by their salinity and temperature, which are conservative properties everywhere but at the surface and in some areas of the seafloor. Nonconservative properties such as oxygen concentration, as well as concentrations of human-made radionuclides and persistent organic compounds, are also used as water-mass tracers.

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8.16.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- back-scattered
- brines
- chaotic
- chlorofluorocarbons (CFCs)
- coastal downwelling
- coastal upwelling
- coastline
- communities
- contours
- convection
- convection cells
- convergence
- Coriolis effect
- cum sole
- currents
- density
- diffusion
- divergence
- downwelled
- dynamic height
- eastern boundary currents
- eddies
- Ekman transport
- El Niño
- fossil fuel
- friction
- front
- geostrophic
- gravity
- gyres
- half-life
- hydrothermal vents
- ice age
- in situ

- internal waves
- intertropical convergence zone
- kinetic energy
- Langmuir circulation
- latitudes
- meanders
- mixed layer
- nutrients
- ozone
- phytoplankton
- phytotrophic
- potential energy
- pressure gradient
- radioactive isotope
- radionuclides

- salinity
- sediment
- shore
- species
- stratified
- subtropical gyres
- surface tension
- suspended sediments
- synoptic
- thermohaline circulation
- tides
- topography
- trade wind
- upwelling
- wavelengths
- weather
- westerly
- western boundary currents

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8.16.2: Chapter Questions

Study Question

1. What are the major energy sources for the surface currents and the thermohaline circulation of the oceans?
2. Why doesn't a wind-driven current flow in the same direction as the wind that causes it?
3. Wind-driven Ekman transport moves surface water across the oceans. How does this lead to geostrophic currents that continue to flow even if the winds abate? Why don't geostrophic currents flow in the same direction as the Ekman transport?
4. Subtropical gyres are present in each ocean and each hemisphere in the latitudes between the trades and westerlies. Why? Would they exist if the Earth were totally covered by oceans? Why or why not?
5. Why do the subtropical gyres rotate clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere?
6. How are upwelling and downwelling caused in the oceans, and why are they important? Where would you expect to find upwelling and downwelling areas in the oceans? Why?
7. What factors affect coastal currents but not deep-ocean currents?
8. Why are the deepest water masses in the oceans formed at high latitudes? Why are they not formed in the Indian and North Pacific oceans?
9. Why is there no pycnocline in most high-latitude regions of the oceans?
10. Why are temperature and salinity the two most important tracers of water masses?

Critical Thinking Questions

1. You might have heard it said that because of the Coriolis deflection, water going down a bathtub drain will swirl in opposite directions depending on whether the drain is located in the Northern or Southern Hemisphere. Do you think this is true? Explain why or why not.
2. Imagine that the viscosity of water was much higher, so that it flowed more like syrup. (a) Would the Ekman spiral and Ekman transport be different? If so, how would they be different and why? (b) Would there still be geostrophic currents? If so, how would they be different?
3. If seawater had a maximum density at 4°C, how would the deep water layers be different from those of today? Where would deep water masses be formed?
4. If we were to build a solid causeway across the Pacific Ocean from Seattle to Tokyo, how would the surface currents in the North Pacific Ocean be affected, and how might this affect the climate of Japan, Alaska, and California?
5. If the Earth were covered in water except for the North American continent, would there be geostrophic gyres? If so, where? What would the current pattern be in the Southern Hemisphere?
6. Discuss why you would expect the Gulf Stream to slow down and Europe to be colder if the rate of formation of North Atlantic Deep Water was reduced.
7. Why is there no geostrophic gyre in the Mediterranean Sea?
8. Global climate may change in the next few decades, with the average temperature of the ocean surface waters increasing by as much as 2–3°C. (a) Hypothesize what might happen to the thermohaline circulation of the oceans as a result. (b) Speculate on what would be the effects on the intensity, geographic distribution, and frequency of upwelling of water from below the permanent thermocline. Explain your reasons for these speculative changes.
9. From what you have learned about ocean circulation, what do you think would be likely to happen to the atmospheric concentrations of carbon dioxide if the carbonate compensation depth (CCD) were to migrate to shallower depths? How long would you expect it to take for these changes to occur? Explain why.
10. If the hypothesized future global warming does indeed occur, the Arctic ice pack may melt completely. (a) How would this affect the bottom water mass of the Arctic Ocean? (b) How might it also affect the characteristics of bottom water masses in the other oceans? (c) What other changes in ocean circulation might be caused? (d) Would these changes be significant during a normal human's lifetime, or would they take thousands of years to occur? Explain the reasons for your answer.

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8.17: Box 1- Westward Intensification

Imagine a motionless wheel suspended horizontally above the Earth's surface at the equator (**Fig. 8B1-1**). To an observer on the Earth, the wheel does not appear to rotate as the Earth rotates. Now imagine that we transport this still motionless wheel to the North Pole and again suspend it horizontally above the Earth. To an observer on Earth, the stationary wheel appears to spin at the same rate as the Earth. This effect also applies to spinning gyres of water in the oceans. The rate of apparent rotation increases with distance from the equator and is called planetary vorticity. Western boundary currents undergo an increase in their apparent rate of rotation as they flow poleward, whereas eastern boundary currents undergo a decrease in their apparent rate of rotation as they flow toward the equator. This is the underlying reason for the westward intensification, but it requires some explanation.

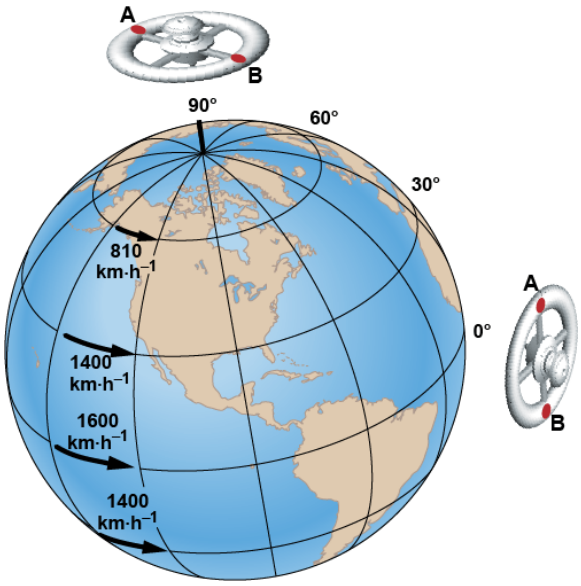

Figure 8B-1. Vorticity (apparent rotation) increases with increasing latitude. Imagine a rotating Earth with a circular space station orbiting above a specific location on the equator but not rotating around its own axis. An observer looking up from the equator would see that the nonrotating space station stays in the same position in the sky and does not rotate. Now imagine the space station being positioned directly over the North Pole, but still not rotating about its own axis. To an observer looking up from the pole, the nonrotating space station would appear to rotate once every 24 h. It is more difficult to explain what an observer would see if the space station were moved from the equator to mid-latitudes. However, the observer would see it rotate, and the rotation rate would increase progressively faster with increasing latitude until, at the poles, it reached one apparent rotation per 24 h.

In a spinning fluid, the angular momentum must be conserved. It is not necessary to fully understand this, but you can see an application of this principle when an ice-skater spins. With the arms close to the body, the skater spins quickly. If the arms are extended, the spin becomes much slower, but momentum is not lost. When the arms are once more brought in, the spin becomes faster again. Of course, friction with the ice does eventually steal momentum, slowing down the spin. Angular momentum and vorticity are related, and vorticity is conserved in much the same way as angular momentum. However, the vorticity that must be conserved is a combination of the planetary vorticity and the relative vorticity. The relative vorticity is the vorticity due to the wind-driven motion of the water in the gyre. The math can be a little complicated, so the process is simplified and summarized in **Figure 8B1-2**.

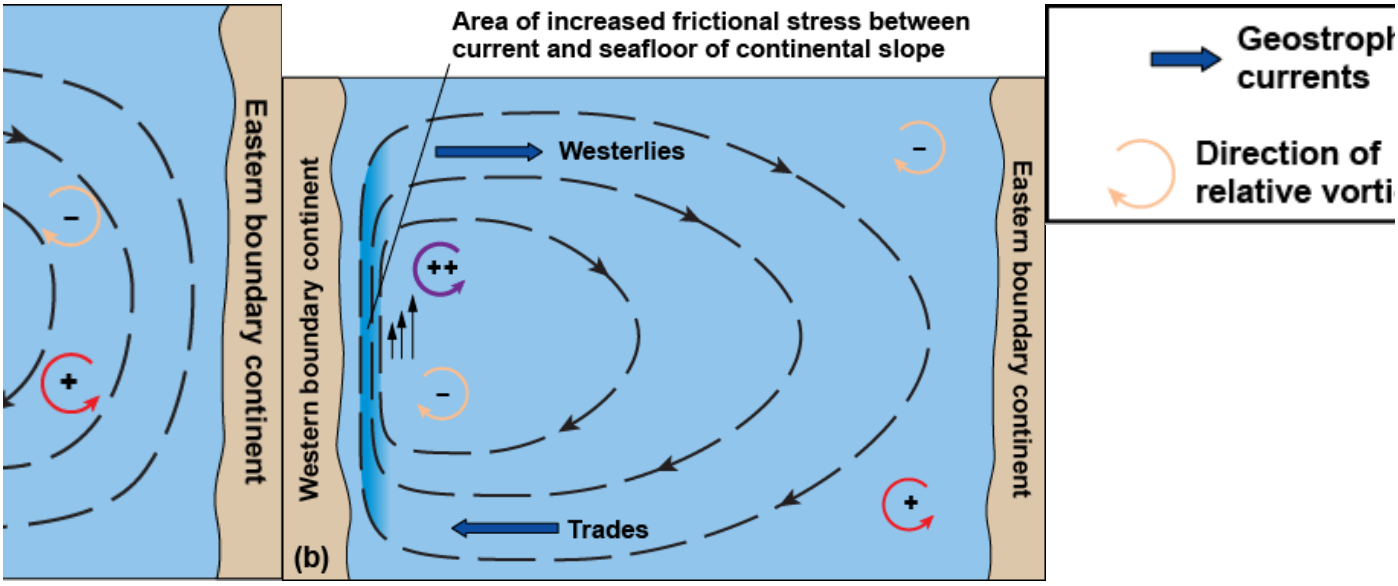

Figure 8B1-2. In the Northern Hemisphere, winds generate clockwise-rotating gyres. This rotation creates clockwise vorticity (or spin) relative to the Earth, and this is called relative vorticity. In addition, vorticity, called planetary vorticity, is imparted by the variations in the magnitude of the Coriolis effect. On the eastward side of the gyre, water flowing southward is subject to a decreasing Coriolis effect, which induces an anticlockwise planetary vorticity (for simplicity, a reduced tendency to turn to the right or spin). On the westward side of the gyre, the northward-flowing water is subject to increasing Coriolis effect, which induces clockwise planetary vorticity. However, the total vorticity, or spin, must be the same for all parts of the gyre. If it were not, then the total mass transport rate of one part of the gyre would speed up while another slowed down, which is not possible. (a) If the gyre were symmetrical, the relative vorticity on the eastward side would be balanced by the planetary vorticity, as they are opposite. On the westward side, planetary and relative vorticity are in the same direction and cannot be balanced. (b) If the gyre is offset (intensified) to the west, friction is increased between the flowing water and the seafloor of the continental slope against which the gyre current is concentrated. The friction provides a source of counterclockwise vorticity to balance the clockwise relative and planetary vorticities.

Figure 8B1-2a represents a symmetrical gyre with no westward intensification. On the east side of the gyre, the planetary vorticity is counterclockwise, and the relative vorticity is clockwise. Thus, the decrease in planetary vorticity as water moves south can be compensated for by an increase in the relative vorticity (speeding up the wind-driven rotation of the gyre, which is in fact what happens as the current flows farther south into the trade wind belt). However, on the west side, both the planetary vorticity and the relative vorticity are counterclockwise. The increase of planetary vorticity could be partly offset if the relative vorticity decreased, but the magnitude of the planetary vorticity at the north end of the western boundary current is much larger than the relative vorticity, so the clockwise motion would be stopped unless there was a strong source of positive (counterclockwise) relative vorticity. This can be supplied by a strong frictional stress that opposes the northward flow of the western boundary current. A strong frictional stress, in turn, can be established if the current is constrained to flow against the land boundary (**Fig. 8B1-2b**), which generates friction at the interface between the current and the seafloor. If the current flows fast and is narrow and deep, the area of contact with the seafloor (along the edge of the continental shelf) is maximized, and, therefore, so is the shear "frictional" stress. As a result, the westward intensification occurs in all ocean current gyres and is caused by the variation in magnitude of the Coriolis effect with latitude.

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8.18: Box 2 - Equatorial Currents

Ekman Transport in Equatorial Region

Because the atmospheric intertropical convergence zone is displaced north of the equator, part of the southeast trade wind zone extends into the Northern Hemisphere. Consequently, although the Coriolis effect is weak near the equator, trade winds are deflected to the right after they pass north of the equator. Once the Southern Hemisphere trade winds have crossed the equator into the Northern Hemisphere, they generate Ekman transport that is deflected to the right, away from the equator, in a generally northeast direction toward the Doldrums (**Fig. 8B2-1**). This pattern contrasts with the Ekman transport to the southwest, or away from the Doldrums, that was assumed in the simple model (**Fig. 8-8**). Ekman transport is directed to the southwest in the major portion of the southeast trade wind belt that lies south of the equator (**Fig. 8B2-1a**).

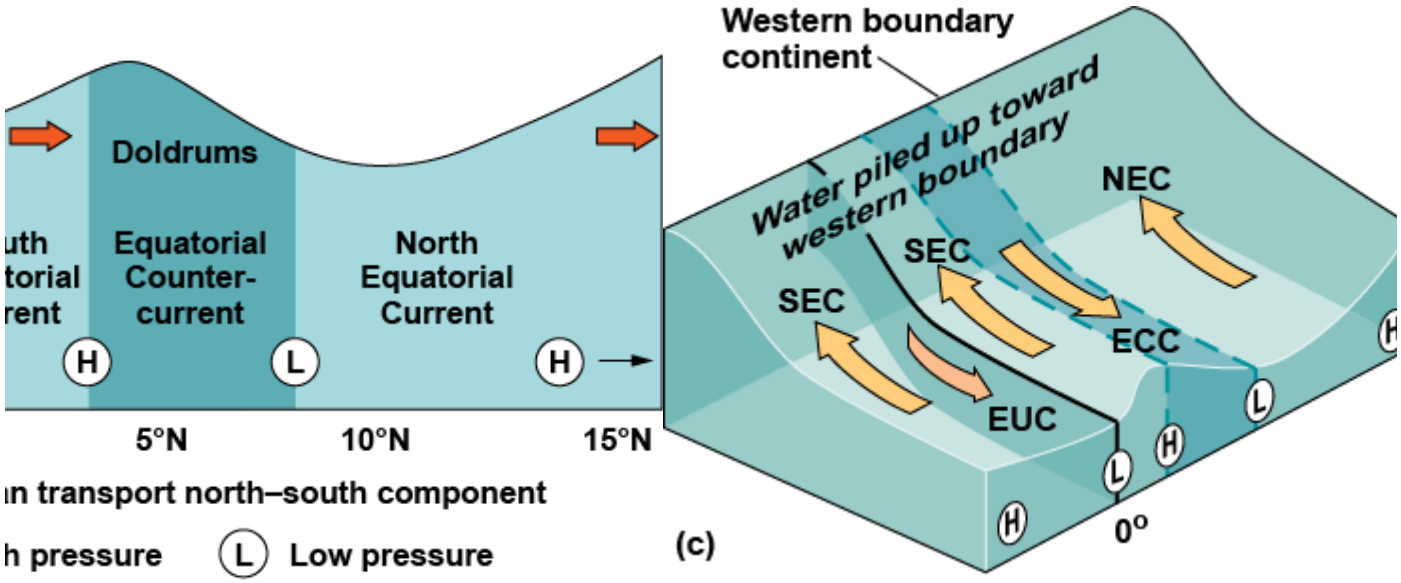

Figure 8B2-1. Currents of the Pacific Ocean equatorial region. (a) The intertropical convergence is displaced north of the equator, especially in the winter, such that the Southern Hemisphere southeast trade winds extend north of the equator. Because Ekman transport is to the right of the wind direction in the Northern Hemisphere, the pattern of Ekman transport just north of the equator is complex. (b) This pattern of Ekman transport produces complex latitudinal variation of sea surface elevation in the equatorial region. (c) The complex sea surface elevation pattern sets up an eastward-flowing Equatorial Undercurrent between the South Equatorial Current and the Equatorial Countercurrent. (d) The Equatorial Undercurrent is a fast current—much faster than the North and South Equatorial Currents or the Equatorial Countercurrent. The Equatorial Undercurrent flows between the surface and a depth of about 300 m, and its maximum velocity occurs at about 100 m.

When the intertropical convergence zone is displaced north of the equator, the sea surface slopes up in a northward direction toward the center of the Northern Hemisphere subtropical gyres throughout the northeast trade wind zone (**Fig. 8B2-1b**). In addition, the sea surface slopes up in a southward direction toward the center of the Southern Hemisphere subtropical gyre throughout the portion of the southeast trade wind zone in the Southern Hemisphere. These slopes are the same as the slopes developed in the simple model of the subtropical gyres shown in **Figure 8-8** and **Figure 8-9**. However, the sea surface slopes upward toward the north in the portion of the southeast trade wind zone that lies between the equator and the Doldrums (**Fig. 8B2-1b**). In this region, Ekman transport is northeast, to the right of the wind, toward the Doldrums, where the sea surface level is undisturbed by local winds.

Geostrophic Currents in the Equatorial Region

As a result of Ekman transport processes, a small “hill” of water is present in the Northern Hemisphere, and its crest is at the intersection of the Doldrums with the southeast trade wind zone (**Fig. 8B2-1b**). In the pressure gradient south of the crest, geostrophic flow is to the west, in the same direction as the geostrophic flow immediately south of the equator. North of the crest, in the pressure gradient within the Doldrums, geostrophic flow is to the east. This is the Equatorial Countercurrent. The arrangement of sea surface slope, pressure gradients, and geostrophic currents in the equatorial region is shown in **Figure**

8B2-1c. The characteristics of each of the currents are described in Table 8B2-1.

Table 8B2-1. *Boundary Currents*

Properties	Western Boundary Currents	Eastern Boundary Currents
Northern Hemisphere examples	Gulf Stream Kuroshio Current	California Current Canary Current
Southern Hemisphere examples	Agulhas Current Brazil Current	Peru Current Benguela Current
Width	Narrow (≤100 km)	Broad (≈1000 km)
Depth	Deep (to 2 km)	Shallow (≤500 m)
Speed	Fast (>100 km•day^{-1})	Slow (<50 km•day^{-1})
Volume transport	Large (50•106 m^{3}•s^{-1})	Small (10–15• ×10^{6})
Boundaries with coastal currents	Sharply defined	Diffuse
Upwelling	Almost none	Frequent
Nutrients	Depleted	Enhanced by upwelling
Fishery	Usually poor	Usually good
Water temperature	Warm	Cool

The equatorial currents of the subtropical gyres and the westward component of Ekman transport in the trade wind zones move very large volumes of surface layer water toward the west. Consequently, the sea surface slopes upward toward the western boundary. Much of the water that flows westward is either diverted north or south in the subtropical gyres or transported back toward the east in the Equatorial Countercurrents. However, water also flows back to the east in a current called the Equatorial Undercurrent.

Equatorial Undercurrents flow at the equator, are about 400 km wide, and have a vertical thickness of about 200 m (Fig. 8B2-1d). In the Pacific, the core of the current rises from a depth of about 200 m in the west to about 40 m in the east. In the Atlantic, the core is at about 100 m depth. Equatorial Undercurrents are very swift, with speeds comparable to those of the Gulf Stream. The Equatorial Undercurrent flows on the west-to-east pressure gradient created by the “piling up” of surface water toward the western boundary. However, the fact that it flows directly down the pressure gradient distinguishes it from other geostrophic currents,

which all flow parallel to the contours of constant pressure (across the pressure gradient) to balance the pressure gradient and the Coriolis effect (**CC12, CC13**).

The Equatorial Undercurrent can flow down the pressure gradient because it flows precisely west to east at the equator, where the Coriolis deflection is zero. If the current deviates slightly to the north, the Coriolis deflection (to the right in the Northern Hemisphere), although weak, turns the current back to the south. If it deviates slightly to the south, the Coriolis deflection (to the left in the Southern Hemisphere) turns it back to the north. Thus, the Equatorial Undercurrent flows directly across the oceans for thousands of kilometers without deviating from its west-to-east direction. The Pacific Ocean Equatorial Undercurrent is particularly important for the critical role it plays in El Niño (**Chap. 7**).

References

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