CHAPTER 7

General Circulation and Secondary Circulations

Chapter at a Glance

Processes occurring within the various components of the climate system have been described in previous chapters. Collectively, these processes contribute to the "real-world" atmospheric circulation and its variations. This chapter explains **general circulation**—the overall, prevailing pattern of winds on large spatial and temporal scales. Certainly, some circulations are local and unusual compared with the overriding, prevailing patterns. At broad spatial scales over a long period of time, however, the atmosphere's preferred directions of circulation are particularly noticeable. These patterns steer energy, matter (especially water and solid *aerosols*), and *momentum* from one part of Earth to another.

Circulation results from fulfillment of the *second law of thermodynamics*. One form of this law states that energy tends to be moved from areas of greater concentration to areas of lesser concentration. Because the Sun heats Earth's surface unequally, a circulation must exist in the atmosphere and oceans that attempts to equalize this energy imbalance. But the energy can never be balanced perfectly because as the general circulation of the atmosphere and the oceans continuously attempts to redistribute the energy, the Sun continuously warms different parts of the earth at different rates.

Specifically, the general circulation redistributes heat (energy) that arrives at the earth in greater quantities near the equator than near the poles. Factors such as the unequal distribution of land mass with *latitude* and variation of landforms and land cover types across the terrestrial earth have concomitant implications for the radiation balance, thereby complicating the general circulation. Chapters 8 through 10 describe the specific climate types that result from this general circulation.

In addition to general circulation, this chapter focuses on **secondary circulations**—smaller circulation systems that are characterized by traveling high- and low-pressure systems that affect climate. These circulations are influenced in the midlatitudes by their position relative to planetary-scale waves of motion. This chapter also examines the relationship between these waves and surface *anticyclones* and *cyclones*.

Circulation of a Nonrotating Earth

Before we can understand the atmosphere's general circulation on the "real" Earth, let us first consider the properties of the circulation if Earth did not rotate. The British meteorologist Sir George Hadley first postulated the general atmospheric circulation in 1735. His studies centered on a simple model of hemispheric *convection*. He posited that air should warm, become less dense, and rise in the area near the equator (**Figure 7.1**). Following the model of a convection cell, he thought that this rising equatorial air would then travel poleward far above the surface (in both the northern and southern hemispheres), ultimately cooling and sinking in the polar areas. This air stream would then diverge at the surface in the vicinity of the poles and move back equatorward across the surface to begin the process anew. The sinking air at the poles from this circulation would produce a permanent surface anticyclone at each pole. A belt of low pressure would exist in the equatorial region, with the lowest pressure along the *longitude* that was being heated most directly—experiencing *solar noon*—at that particular time of day.

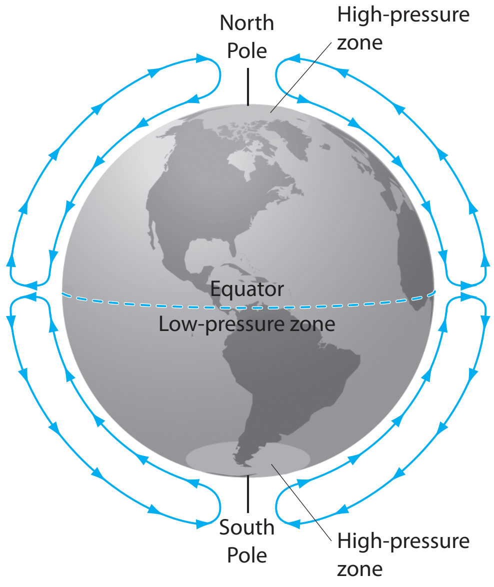


Figure 7.1 The simple hemispheric convection cell model of atmospheric circulation proposed by Sir George Hadley, a British meteorologist.

The major problem with Hadley's view of atmospheric circulation was that it did not match observed data of wind patterns reported from sailing vessels across the globe. The circulation of the planet is more complex than the simple single hemispheric convection cell pattern suggested by Hadley. Still, Hadley's work was the first to include the upper atmosphere in thoughts centered on the general circulation of the atmosphere and for that it gains merit.

There are two main complications that Hadley did not consider. The first is that Earth's surface is not uniform. Rather, it has many different topographic and land/water surface variations that upset the formation of a single hemispheric convection cell in each hemisphere. Because these variations are often local or regional, they tend to contribute even more to local and synoptic circulation patterns than to changes in the "general" circulation. The second complication—the rotation of Earth—exerts a more direct influence on the general circulation. This rotation initiates trajectory changes in moving fluids through the *Coriolis effect*.

Idealized General Circulation on a Rotating Planet

Hadley postulated correctly that, in general, air near the equator gains thermal buoyancy because of its warmth. He was also correct to infer that the cold air near the poles has a tendency to sink. We have already seen that rising air is generally associated with atmospheric low-pressure situations, whereas sinking air is typically linked to high-pressure systems. Thus, a broad belt of low surface pressure dominates the equatorial region, whereas higher surface pressures exist over the poles.

Hadley Cells

Hadley also proposed that air rising near the equator begins moving toward the poles in the upper *troposphere*. This upper-level poleward motion occurs because of an upper atmospheric pressure gradient that pushes air away from the equator (where there is an excess of pressure aloft caused by the "extra" air that rose from the surface). **Figure 7.2** shows a model of the general circulation on a rotating planet.

Unlike the model originally proposed by Hadley, most of the air aloft flowing away from the

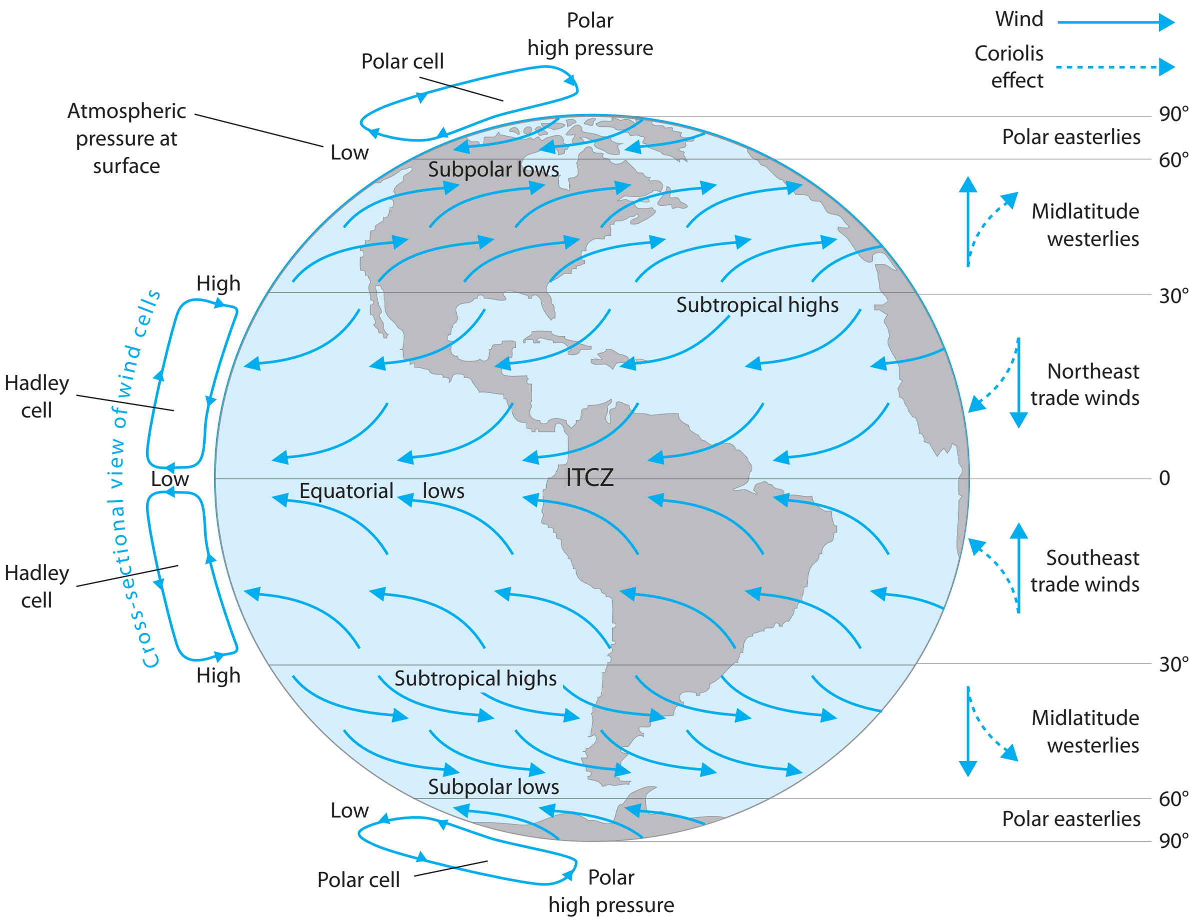


Figure 7.2 Pressure and wind belts of the generalized three-celled general circulation model.

equator does not actually travel all the way to the poles. Instead, much of it cools while it is high above the surface and sinks in the vicinity of 30° latitude (both north and south of the equator). This air diverges from the point of impact at the surface near 30° latitude, forming the *subtropical anticyclones*, or subtropical highs, which are centered at approximately 30° latitude in each hemisphere. As we see later, the highs are situated over each major ocean basin.

The region of the subtropical anticyclones is also known as the **horse latitudes**—a term derived from the days of European exploration when sailors are said to have discarded horses overboard when they found themselves under the stagnant regions beneath the high-pressure zones. The weak descent of air through the atmosphere in these regions provides little in the way of steering currents necessary to push sailboats across vast distances. Thus, sailing vessels frequently stalled in these areas. Water shortages ensued, which sealed the fate of horses onboard. The subtropical anticyclone region of the North Atlantic Ocean where the **Bermuda-Azores high** is located is also known as the **Sargasso Sea** because of the prevalence of the sargasso seaweed that accumulates seasonally in the rather calm, stagnant waters under sunny skies.

This circulation between the equator and 30° latitude is largely convective, as vertical motions within it are driven solely by heat energy. This convection cell, with rising motion near the equator and sinking motion near 30° latitude, is termed a **Hadley cell**; one Hadley cell exists in each hemisphere.

Polar Cells

Near the poles the net energy deficits reinforced by the high *albedo* of snow and ice in polar areas cause the air to increase in *density*, which initiates sinking motions through the atmosphere. This air diverges at the surface, inducing the permanent high atmospheric pressure at the surface—the **polar highs** (one at each pole). Upon reaching the surface, the diverging air travels equatorward toward lower latitudes, is deflected by the Coriolis effect, and eventually converges at approximately 60° latitude with air that was diverging from the subtropical anticyclones at 30°. The converging air near 60° latitude is forced to rise. As we have seen, rising air motions are associated with low atmospheric pressures. The convergence creates **subpolar lows**—areas of relatively low surface pressure centered on 60° latitude in each hemisphere. These circulation systems are the **polar cells;** one polar cell exists in each hemisphere.

Planetary Wind Systems

As we have seen, starting at the poles and moving toward the equator at 30° intervals of latitude are surface oscillations of high (at 90°), low (at 60°), high (at 30°), and low (at 0°) pressure in each hemisphere (Figure 7.2). These differences in pressure across space set up differences in atmospheric mass, because pressure is a function of force (which is the product of mass and acceleration according to *Newton's second law*). Because mass is transferred from areas of higher concentration to areas of lower concentration, this general surface pressure distribution sets up the planetary wind systems. Winds are simply the transfer of atmospheric mass between an area of excess mass and an area containing less mass. Because pressure differences define winds, let's begin our discussion with an examination of the influence of the major pressure features on surface wind patterns.

Surface Wind Systems As with all surface anticyclones, air in the subtropical anticyclones at approximately 30° latitude in each hemisphere sinks and diverges across the surface in all directions. The *pressure gradient force* pushes air directly toward lower pressure areas—namely, toward 0° and 60° latitude in each hemisphere, the equatorial and subpolar lows, respectively. Because of the rotation of the planet, the Coriolis effect deflects these surface winds to the right in the northern hemisphere. This causes air diverging from the northern hemisphere's subtropical highs to flow clockwise (anticyclonically) as it moves outward. Likewise, air diverging from the southern hemisphere's subtropical highs flows counterclockwise—also anticyclonically for that hemisphere—as it moves outward away from the high, as the Coriolis effect deflects motion to the left in that hemisphere (Figure 7.2).

Thus, the surface winds moving between 30°N and the equator take on a motion that is from the northeast. As we saw in Chapter 3, winds are always named for the direction from which they blow. These winds are, therefore, appropriately called the *northeast trade winds* (or northeast trades). They are termed "trade winds" because they constituted the primary sailing trade route between southern Europe and Spanish America across the North Atlantic Ocean. The northeast trade winds steered Christopher Columbus from his point of departure in Spain toward his unintended destinations in the Caribbean, Central America, and northern South America. A strong argument can be made that Spanish is the primary language spoken in Latin America and the Caribbean because of the northeast trade winds!

The southern hemisphere also has trade winds, but the leftward deflection of the Coriolis effect causes air flowing out of the subtropical anticyclones to move counterclockwise and surface winds to move primarily from the southeast between 30°S latitude and the equator. These are called the *southeast trade winds* (or southeast trades). It should be easy to remember that the northeast trades occur in the northern hemisphere, whereas the southeast trade winds occur in the southern hemisphere.

Because the northeast and southeast trade winds converge into the low-pressure equatorial low, the equatorial low is sometimes known as the **Intertropical Convergence Zone (ITCZ)**. Although the name appears rather imposing, it is quite explanatory as it describes surface airflow across the tropics that converges in a zone across the low latitudes. The ITCZ also carries another name—the **doldrums**, a belt of relatively stagnant winds that form as air rises from the surface because of trade wind convergence from both hemispheres. Sailing ships venturing into this area would often stagnate for long periods. Being "stuck in the doldrums" achieved colloquial status, signifying a depression of some sort. Interestingly, the doldrums became the region of many mutinies, including the famous mutiny of Captain Bligh on the *HMS Bounty*.

On the poleward sides of the subtropical anticyclones, the surface pressure gradient force pushes air from the high-pressure core toward the subpolar lows. This air is then deflected to the right (in the northern hemisphere) or the left (in the southern hemisphere) on its trek to fill in the subpolar low. In this case, though, the latitude is higher than in the case of the tropical trade winds, so the Coriolis deflection is stronger. Recall that the magnitude of the Coriolis deflection is proportional to latitude. This strong deflection causes the general circulation between 30°N and 60°N latitude to be from west to east near the surface. This surface wind belt is known as the **midlatitude westerlies**. The southern hemisphere also has westerlies between 30°S and 60°S latitude. Even though the Coriolis deflection in that hemisphere is from the opposite direction, the fact that the midlatitude zone is south of the subtropical anticyclone rather than north of the high (as in the northern hemisphere) causes surface westerly winds to exist in the southern hemisphere as well.

Air sinking to the surface and diverging equatorward at the polar highs also undergoes Coriolis deflection. As the air travels from the region of the North Pole toward lower latitudes, it is deflected strongly (because of the high latitude) to the right. This creates the surface wind belt known as the **polar easterlies**, as near-surface prevailing air is generally from the east. Surface air moving from the region of the South Pole is deflected to the left as it moves toward that hemisphere's subpolar low. Coriolis deflection to the left also creates polar easterlies near the surface for that hemisphere between 90° and 60°S latitude.

To summarize, moving from the region of the poles toward the equator, the following surface wind belts occur:

- Polar easterlies (90–60°N latitude)
- Midlatitude westerlies (60–30°N)
- Northeast trades (30°N–0°)
- Southeast trades (0–30°S)
- Midlatitude westerlies (30–60°S)
- Polar easterlies (60–90°S)

Upper-Level Winds in the Hadley and Polar Cells Upper-level winds overlaying the surface wind patterns are also an important component of the general circulation. As is shown in Figure 7.2, the upper-level pressure gradient force pushes air in the opposite direction from that at the surface in both the Hadley cells and the polar cells. At the surface the pressure gradient force is directed from the subtropical anticyclones to the equatorial low in the Hadley cells. But aloft, the upper-level pressure gradient force pushes air from the equatorial region toward the poles, therefore, moving air over the subtropical anticyclones. Over the polar cell the upper-level pressure gradient force pushes the air from the area over the subpolar lows to the area over the polar highs, a direction opposite to that at the surface. Of course, in both the Hadley cells and the polar cells, the Coriolis effect causes a deflection of these winds generated by the pressure gradient force. The deflection is less in the Hadley cell than in the polar cell because of the lower latitudes of the Hadley cell. The result is upper-level southwesterlies over the northern hemisphere Hadley cell, northwesterlies over the southern hemisphere Hadley cell, and westerlies over the polar cells in both hemispheres (**Figure 7.3**). These polar westerlies aloft advected the tephra from the Eyjafjallajökull volcanic eruptions in Iceland eastward toward Europe beginning in the spring of 2010.

As air aloft in the Hadley cell continues to move poleward (the southwesterlies aloft in the northern hemisphere and the northwesterlies aloft in the southern hemisphere), the Coriolis effect increases in strength. The deflection to the right of the flow (or to the left of the flow in the southern hemisphere) increases poleward, and flow becomes increasingly westerly near the poleward edge of the Hadley cell in each hemisphere. Winds at these positions are often very strong, in large part because of **conservation of angular momentum**—an important property of curved motion in a fluid on Earth. Anything that is "conserved"—in this case, angular momentum—remains constant.

Like any type of momentum, angular momentum is the product of a mass and a velocity. Angular momentum simply includes an additional variable—the radius of curvature. The "tighter" the curve, the smaller is the radius of curvature. An infinitely large radius of curvature occurs for straight-line flow. The conservation of angular momentum can be expressed quantitatively as

$$mvr = \text{constant}$$

where m is mass, v represents velocity, and r is the radius of curvature. In the atmosphere mass is virtually constant. Therefore, the equation above

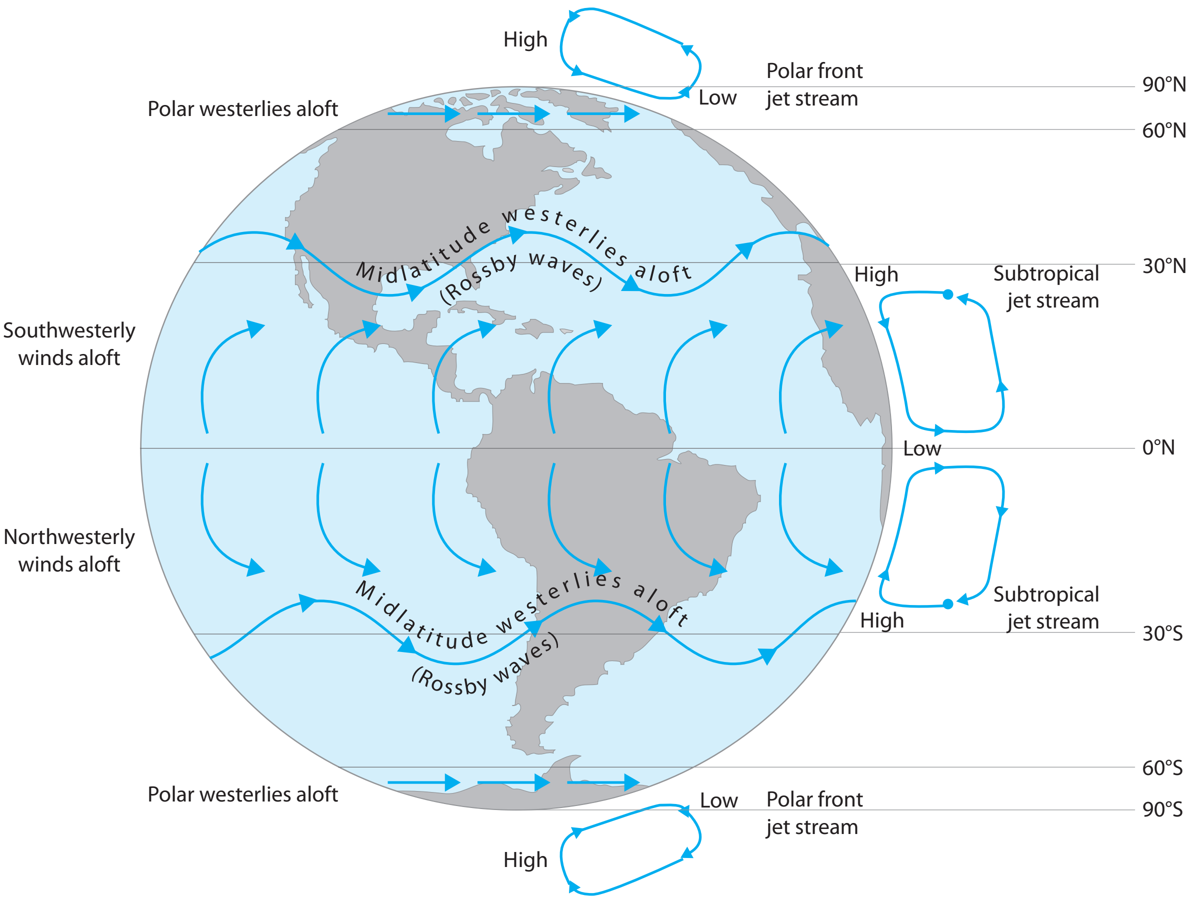


Figure 7.3 Global upper-level wind systems.

suggests that as *r* decreases, *v* increases in order to maintain a constant angular momentum. Ice skaters know that to spin faster (increase velocity), they must become "thinner" by moving their arms inward to reduce their radius of curvature. In the atmosphere, westerly winds moving around the hemisphere at higher latitudes have smaller radii of curvature (the distance between the westerly wind and the north Earth's axis) than westerly winds moving around the hemisphere at lower latitudes. As air moves poleward, its radius of curvature decreases (i.e., it goes from the wider part of Earth near the equator to the narrower part of Earth near the poles) and begins to curve increasingly to the right (in the northern hemisphere) because of the increasing Coriolis effect. Its velocity must increase simultaneously as *r* decreases to ensure that angular momentum is conserved. The result is the westerly *subtropical jet stream* that exists near 30°N and 30°S latitude (Figure 7.3). Although much remains to be discovered about the influence, variability, and impact of this fast-moving river of air in the upper troposphere, it is known to exert a significant impact on weather and climate in subtropical latitudes.

Upper-Level Winds in the Midlatitudes To understand the upper-level winds overlying the midlatitude surface westerlies, upper-level pressure gradients can be considered in terms of **geopotential height**, which is simply the altitude (in meters above sea level) of a given pressure surface over a particular location at a certain time. For example, the 500-mb geopotential height represents the altitude at which 500 mb of pressure is being exerted by the weight of the atmosphere in an atmospheric column. Because mean sea level pressure is 1013.25 mb, the 500-mb geopotential height also represents the approximate middle of the atmosphere from a mass perspective; roughly half of the atmospheric mass is above and the other half is below the 500-mb geopotential height level.

In relatively warm air, such as in the equatorial part of Earth, the surplus of heat increases the molecular *kinetic energy* of the gases in the air column. As this energy causes molecules of air to expand upward, the warm air becomes less dense than the colder air as the gases occupy more volume and, therefore, weigh less per unit volume than when pushed closer together near the surface. We would expect that the 500-mb level (and other constant-pressure levels) over the equatorial part of Earth would be relatively high. In a warm column of atmosphere the elevation that divides the total mass of the atmosphere in half is relatively high because the molecules are spread out more vertically in the column. In the polar part of Earth, the high density of cold air occurs because molecular kinetic energy in the individual gas molecules is relatively low. This causes a contraction of the air column and a downward movement of the constant-pressure levels.

In the real world, then, the net energy surplus within the tropics causes tropospheric thermal expansion in that area, and the net energy deficit at the poles causes tropospheric contraction. The equatorial *tropopause* averages approximately 19 km (12 mi) above the surface, but the polar tropopause averages only about 10 km (6 mi) above the surface. A steady decrease in upper-level geopotential heights tends to occur from the equator toward each pole as colder conditions prevail (**Figure 7.4**). Furthermore, the difference in a given geopotential height level between the equator and pole increases with increasing height in the troposphere. For example, it is apparent from Figure 7.4 that the 850-mb geopotential height is somewhat higher over the tropics than over the poles. But above the 850-mb level, the tropics are also warmer than the poles (particularly in winter), so the 700-mb geopotential height difference between the equator and the pole is compounded on the difference that already occurred at the 850-mb level. The latitudinal slope in geopotential heights creates a pressure gradient that pushes air horizontally from areas with higher heights (lower latitudes) to areas with lower geopotential heights (higher latitudes). Because the gradient steepens and friction weakens with increasing height, we expect the winds to strengthen with height. In the Hadley and polar cells, these upper-level winds cannot strengthen as much because they are constrained by the flow in the rest of the cell, including that at the surface.

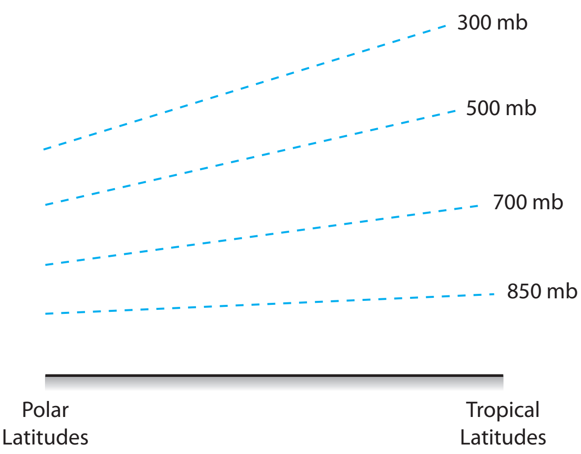


Figure 7.4 The slope of geopotential height surfaces from the equator to the North Pole. A similar slope would occur in the southern hemisphere, as heights decrease toward the South Pole.

When upper-tropospheric air moves poleward in response to this midlatitude height gradient, the Coriolis effect deflects it. Ultimately, a balance is gained between the pressure gradient force and Coriolis effect, which results in upper-level westerly winds in both hemispheres from the tropics all the way to the poles. The midlatitude region is the only one of the three zones per hemisphere (Hadley, polar, and midlatitude) in which upper-level winds occur in the same direction as the mean surface winds.

Modifications to the Idealized General Circulation: Observed Surface Patterns

Land–Water Contrasts

The model described above is a highly idealized conception of the circulation system of Earth. It assumes that there are no surface variations—the model applies only to a planet composed solely of water. But in the real world, land–water contrasts complicate circulation in the climate system. As we saw in Chapter 3, large water bodies maintain relatively constant temperatures through the course of a year compared with inland areas, assuming that all other factors are equal. The consistency is caused by a combination of water transparency, high *evaporation* rates, vertical and horizontal currents, and a high *specific heat* in water, all of which dampen fluctuations in temperature near the water surface.

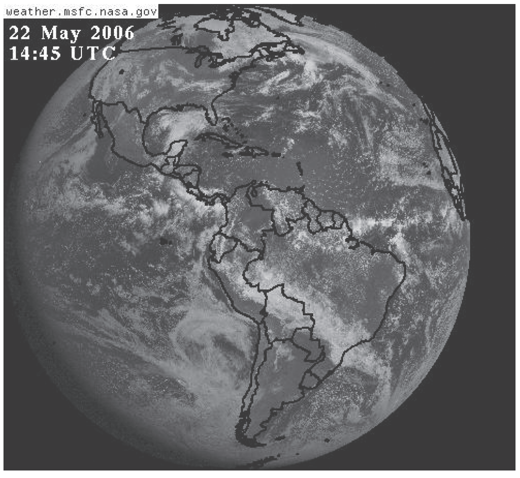
On the other hand, continents show wide temperature variations across the seasons because their opaque surfaces absorb *insolation* in a tiny surface layer and then readily transfer that energy back to the atmosphere. Inland surfaces also have relatively low evaporation rates (causing a relatively high percentage of energy to be devoted to *sensible heat*), virtually no convective/advective motion to redistribute energy, and a low specific heat. These factors produce surfaces that heat quickly when energy is present and cool quickly during periods of energy deficit.

Given these factors and variations in surface elevations associated with continents, it is no surprise that including continents into the idealized model causes disruptions to the circulation features. But the idealized model does provide a useful starting point to understanding the more complex real-world features more fully.

Locations and Strength of Features in the Hadley Cells

ITCZ and Trade Winds The ITCZ exists in a similar form in the real world as in the idealized pattern (**Figure 7.5**). It may be thought of as a constantly moving convective thunderstorm chain that follows the migration of the vertical solar ray across varying longitudes through the course of a day and between the tropical latitudes through the course of the year. But the ITCZ is much more physically constrained by surface temperature characteristics than the faster-moving solar rays. It, therefore, lags considerably behind the solar radiation maximum in its latitudinal migration. Also, it never reaches the maximum latitudinal position of the 90° vertical ray, which is at the *Tropic of Cancer* (on June 21) and the *Tropic of Capricorn* (on December 22) (**Figure 7.6**).

(a)



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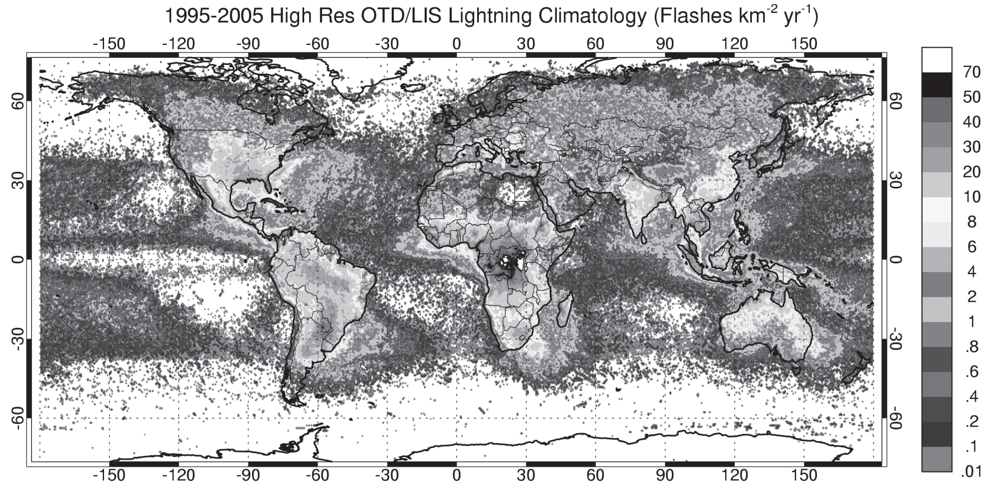

Figure 7.5 Global view of ITCZ represented by (a) the line of low-latitude clouds and (b) the most concentrated lightning flashes. Notice how the ITCZ is most prominent at the longitudes experiencing afternoon at a given time. (See color plate 7.5b).

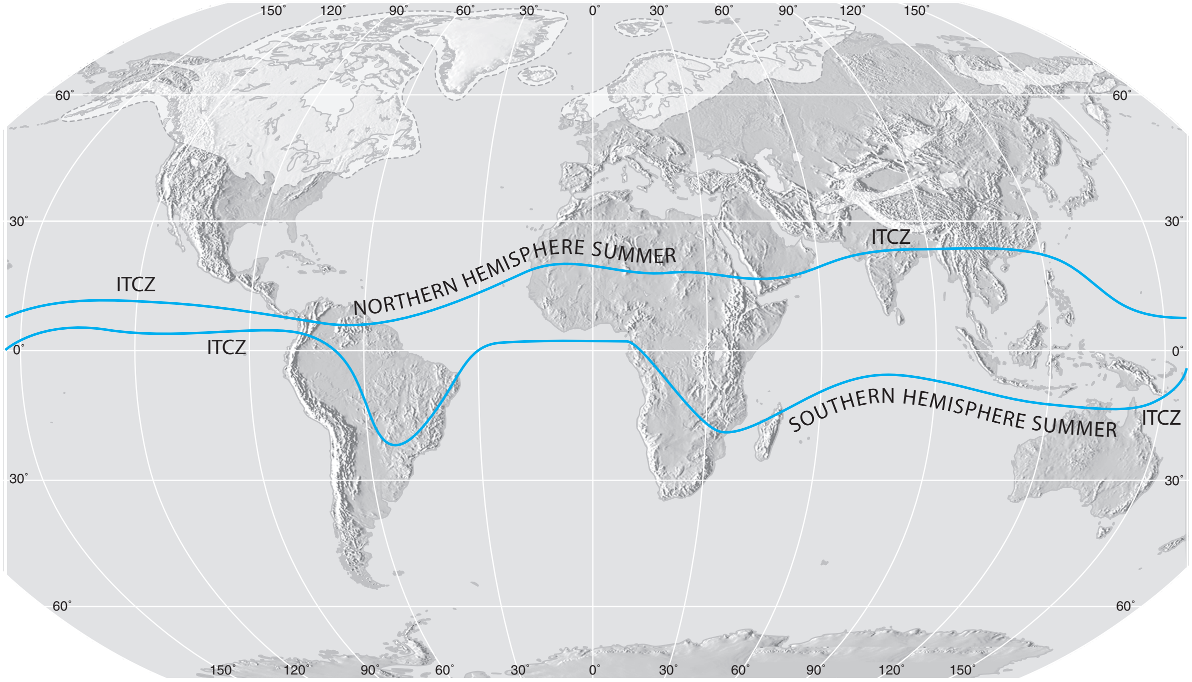

Figure 7.6 Migration of the ITCZ through the year. *Data from*: NOAA/GOES.

Instead, the ITCZ tends to migrate only approximately 10–20° of latitude away from the equator in most locations, lagging behind the *solar declination*. The ITCZ moves poleward into the summer hemisphere in response to the displacement of the vertical solar rays. The ITCZ does not migrate much over places affected by cold ocean currents, such as the tropical west coast of South America, even if these locations are in the path of the Sun's vertical rays.

As the ITCZ migrates seasonally, the trade winds also migrate—to their northernmost extent in northern hemisphere summer and to their southernmost extent in northern hemisphere winter. Thus, some locations on Earth may experience the trade winds for only part of the year. This migration can have a significant influence on the seasonal climate of tropical locations.

Subtropical Anticyclones The idealized model suggests that a continuous belt of surface high pressure should exist across the planet near 30°N and 30°S—the subtropical anticyclones. In reality, however, these anticyclones occur only over the ocean basins near 30° latitude, and they are larger in summer than in winter (**Figure 7.7**). The reason is that the tendency for high pressure at 30°N and 30°S is reinforced by the *relatively* cool summer conditions (compared with the adjacent land surfaces) over the oceans. The relatively cool ocean surface chills the overlying air, increasing low-level atmospheric *stability*. During that time internal pressure in the subtropical anticyclones reaches a maximum. The subtropical anticyclones are strongest when and where the ocean surface is much colder than the continental surface in the subtropics. The sinking air that results diverges at the surface, spilling toward lower-pressure locations in all directions before experiencing Coriolis deflection.

By contrast, the subtropical land surfaces heat effectively under long day lengths and high solar angles in summer. They tend to encourage rising motion (and lower surface pressure) that offsets the general circulation's tendency for sinking motion and high surface pressure. The subtropical continents are, therefore, typically too warm in summer to support the development and maintenance of semipermanent surface high-pressure cells.

The subtropical anticyclones migrate somewhat from their summer position, latitudinally with solar declination and longitudinally depending on where the coldest water lies, within the subtropical oceans. They also weaken in winter because the ocean surface is *relatively* warm compared with the land surfaces at similar latitudes. This relative

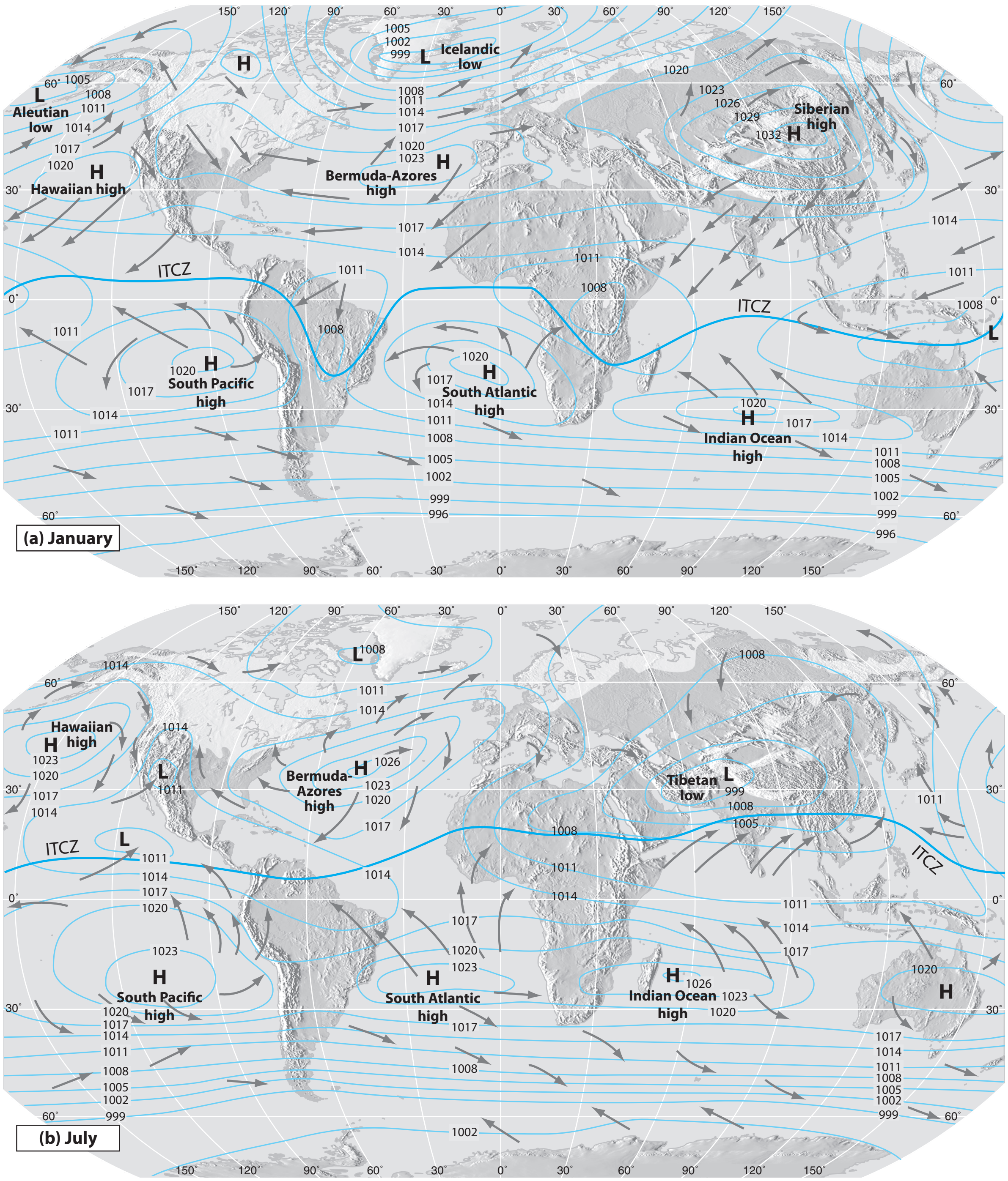

Figure 7.7 Mean surface isobars, positions of the semipermanent surface pressure cells, and associated winds, during (a) January and (b) July. *Adapted from*: January Pressure and Predominant Winds, and July Pressure and Predominant Winds, 1995. *Goode's World Atlas,* 19th ed. Edited by E.B. Espenshade. Skokie, IL: Rand McNally and Co., 1995.

warmth induces rising motion that tends to offset the tendency for sinking air at 30°N and 30°S, as suggested by the idealized general circulation model. Instead, pressure tends to increase over the relatively cold terrestrial atmosphere.

The major surface-based, semipermanent, high-pressure cells located near 30°N latitude are the Bermuda-Azores high in the North Atlantic Ocean and the **Hawaiian high** (also known as the **North Pacific high**) in the North Pacific Ocean

(Figure 7.7). Southern hemisphere subtropical anticyclones include the **South Pacific high**, the **South Atlantic high**, and the **Indian Ocean high**.

Locations and Strength of Features in the Polar Cells

Surface pressure features in the polar cells also strengthen and weaken seasonally. During winter the lack of insolation in the high latitudes causes intense surface chilling, which greatly increases low-level atmospheric stability. The frigid, dense air sinks in the vicinity of the poles and moves equatorward vigorously. The sinking air creates high surface pressure in the vicinity of the poles—the polar highs. These permanent anticyclones—one in each hemisphere—reach their maximum strength during winter and weaken somewhat in summer as the long summer days provide a bit of warming even at the poles. The intensification of the polar high in the winter hemisphere extends the entire polar cell equatorward, including the subpolar low-pressure zones, to approximately 55°N and 55°S latitudes. The weakening of the polar high in the summer hemisphere contributes to the contraction of the polar cell to approximately 65°N and 65°S latitude, coincident with the poleward migration of the subtropical anticyclones.

Like the subtropical anticyclones, the subpolar lows are confined to the ocean basins rather than being a continuous feature across the globe as depicted in the idealized model. The subpolar lows are also similar to the subtropical anticyclones in that their existence is tied to surface temperature variations between land and sea. In winter, oceanic regions at the subpolar latitudes are generally warmer than the adjacent continental locations, both because of the thermal storage properties of water and because of warm ocean currents that transport energy from lower latitudes. The relatively warmer ocean surface promotes low-level atmospheric instability, thereby reinforcing the tendency for rising motion as suggested by the idealized general circulation. This supports the development and maintenance of semipermanent surface cyclonic cells, particularly in winter—the subpolar lows. Buoyant air in the subpolar lows is associated with strong pressure gradients, windiness, and storminess, particularly during winter. The semipermanent North Atlantic subpolar cyclone is known as the **Icelandic low** and that in the North Pacific is referred to as the **Aleutian low** (Figure 7.7). Southern hemisphere subpolar lows are very weak and exist only in a broad, continuous belt.

Continental areas at approximately 55–60°N latitude are frigid in winter. These cold land surfaces support sinking motion, which counteracts the general circulation's tendency for having rising motion at these latitudes. During winter, the *continentality* over the massive Asian continent sometimes causes the lowest surface temperatures for the northern hemisphere to exist over Siberia rather than at the North Pole. This high surface pressure expands over Asia, even into the zone where the idealized general circulation suggests that rising motion should occur. This surface anticyclone is known as the **Siberian high**. In the southern hemisphere, no such continental areas exist at 55–60°S latitude, so there is no equivalent to the Siberian high.

During summer, the long day lengths in high-latitude locations increase continental surface heating substantially, particularly in the northern hemisphere. Under such conditions the thermal gradient between the land masses and the oceans in the vicinity of 60° latitude decreases, weakening the pressure gradient. With *relatively* cooler oceans in summer (compared with adjacent land surfaces), the tendency for rising motion over the subpolar low regions is reduced, and the semipermanent surface oceanic subpolar lows weaken and shrink, as shown in Figure 7.7. At the same time, the Asian land mass heats considerably, once again because of continentality, and the Siberian high disappears. Instead, a semipermanent zone of low pressure occurs over south-central Asia—the **Tibetan low**. The flip-flop of pressure over the Asian land mass between the Siberian high in winter and the Tibetan low in summer as a result of continentality is the primary cause of the monsoon circulation, discussed in Chapter 9.

Regardless of season, the subpolar lows are generally located where the cold surface air moving equatorward from the polar high meets the much warmer poleward-moving air diverging from the subtropical anticyclones. As we have seen, this convergence is displaced poleward in summer and equatorward in winter. The convergence of these two surface air streams causes rising motion that is enhanced by the temperature differences between the two streams, with the warmer tropical air pushed up vertically over the colder polar air. Thus, this region is the boundary between cold, dense, polar air and warm, less dense, tropical air. Storm formation (or **cyclogenesis**) frequently occurs along this band of latitudes because the steep thermal gradient supports the development and motion of migratory cyclones as secondary circulation features.

Locations and Strength of Surface Midlatitude Features

The seasonal fluctuation in strength of the subpolar lows and subtropical anticyclones works in tandem. In the winter hemisphere the subpolar lows are strengthened and displaced equatorward, while the subtropical anticyclones are relatively weak and displaced equatorward also. In the summer hemisphere the subtropical highs are strengthened and displaced poleward, whereas the subpolar lows simultaneously shrink and retreat poleward. The areas between the subpolar lows and the subtropical highs—the midlatitudes—are affected differently throughout the year by these pressure patterns.

Both the subpolar lows and the subtropical highs are affected more by the changing thermal surface characteristics of the northern hemisphere than by those in the southern hemisphere. The increased concentration of land area in the northern hemisphere promotes stronger land–sea contrasts than in the southern hemisphere. Likewise, the semipermanent subtropical highs and subpolar lows are more consistent in their presence and strength in the southern hemisphere because of the lack of large land masses at the subtropical and (especially) subpolar latitudes.

Putting It All Together: Surface Pressure Patterns and Impacts

To summarize, the summer hemisphere sees weakening and contraction of the polar high and subpolar lows and expansion of the subtropical highs. The ITCZ pushes poleward through the tropical part of the summer hemisphere at the same time. This migration of the ITCZ contributes heavily to tropical precipitation regimes. The winter hemisphere sees expansion and strengthening of the polar high and the oceanic subpolar lows into lower latitudes, whereas the subtropical anticyclones weaken and retreat equatorward.

The waxing and waning of these pressure systems affect and are affected by the changing of the seasons in the high and midlatitudes. The changes in the semipermanent circulation cells trigger direct precipitation regime changes for many high and midlatitude locations. In lower latitudes, precipitation changes directly define the seasons because thermal changes are minimal or nonexistent because of day length consistency through the course of the year. There, the ITCZ is the primary precipitation-forcing mechanism. When the ITCZ migrates over or near a region, it brings thunderstorms and associated precipitation. This occurs during the high-sun period of "summer," which is defined not so much by higher temperatures as by direct rays of sunshine falling in the same hemisphere as the point of interest. During the lower Sun season, the ITCZ migrates to the opposite hemisphere, taking with it the primary precipitation-forcing mechanism. This causes a distinct wet/dry seasonal regime for many locations.

Modifications to the Idealized General Circulation: Upper-Level Airflow and Secondary Circulations

As we have seen, flow in the upper-level midlatitudes is primarily westerly. However, these upper-atmospheric air currents typically exhibit flow characteristics that meander northward and southward as they move in the general west-to-east direction. We may characterize the broad-scale, upper-level atmospheric flow at a given time based on the *amplitude* or "waviness" of the flow pattern. If the upper airflow exhibits a relatively deamplified, or nearly straight, west-to-east pattern on a given day, the atmospheric pattern is classified as having a **zonal flow** (**Figure 7.8a**). If significant latitudinal amplification occurs in the midlatitude flow pattern on a given day, then the atmospheric pattern is said to have a **meridional flow** (Figure 7.8b).

Virtually all airflow has at least some meridional component, meaning that the flow pattern is at least somewhat wavy. Because some amplitude is present in all flow regimes, any upper-airflow pattern may be broken into wave segments. Any portion of the wave pattern in either hemisphere

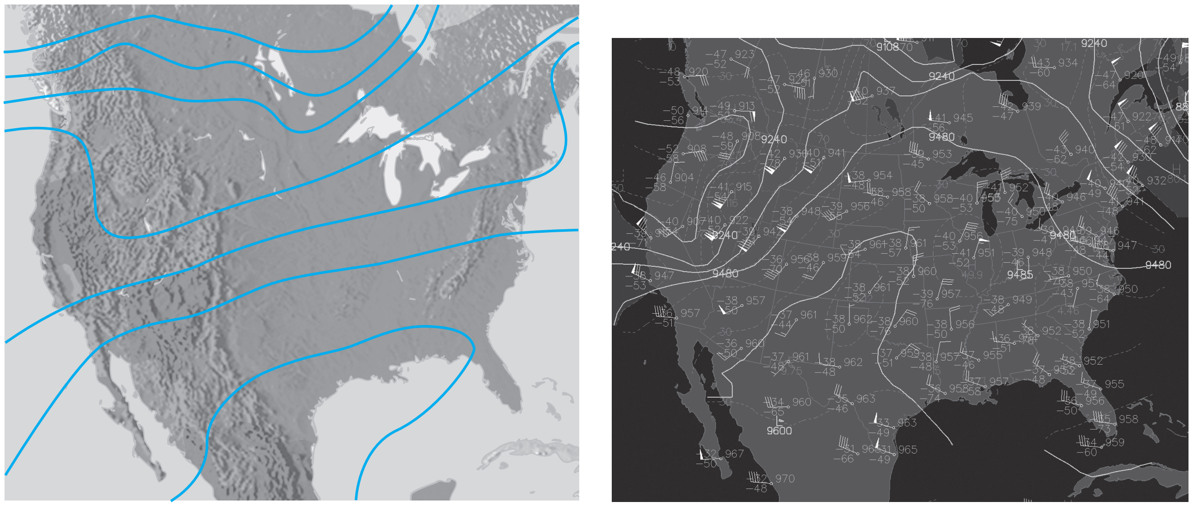

Figure 7.8 (a) The zonal flow over most of the United States, showing a 300-mb geopotential height, recorded at 8 AM EDT on September 18, 2005. *Data from*: The Ohio State University WWW Weather Server, courtesy of the Department of Geography/Atmospheric Sciences Program. (b) A 300-mb map of the United States, recorded on May 28, 2010. On this date, the strongest flow is over the western U.S. and eastern Canada. Reproduced with permission of Unisys Corporation © 2010.

that "bumps" poleward is termed a *ridge*—an elongated zone of high pressure (as opposed to an *enclosed* area of high pressure, which is an anticyclone). The elongation of high pressure refers to the fact that where the midlatitude flow moves poleward, the high pressure of the subtropical anticyclones can be elongated poleward as well. A ridge at the 300-mb level is shown in Figure 7.8b centered over the upper Mississippi Valley.

Any portion of the flow pattern that "dips" equatorward is termed a *trough*—an elongated area of low pressure. By contrast, a cyclone is an *enclosed* zone of low pressure. In Figure 7.8b a 300-mb trough is situated over the western states. When the midlatitude flow dips equatorward, the relatively low pressure associated with the subpolar lows can also be elongated toward the equator.

Interestingly, ridges in the southern hemisphere appear the same as troughs in the northern hemisphere, and vice versa. This is because a dip *southward* of midlatitude flow in the southern hemisphere represents a place where the subtropical anticyclones to the *north* can be elongated, creating a ridge. Likewise, a jog to the north in the southern hemisphere represents a location where the subpolar low-pressure systems can be elongated equatorward, thereby forming a trough.

Like other types of waves, atmospheric waves can be described according to *wavelength*—the distance from one wave crest (ridge) to an adjacent wave crest (or the distance from one wave trough to an adjacent wave trough). Typically, midlatitude waves that have wavelengths of hundreds of kilometers are simply called **long waves**, or **Rossby waves**, named after the atmospheric physicist Carl Rossby, who was instrumental in explaining upper-atmospheric dynamics during the first half of the twentieth century. Rossby waves are continuous around the hemisphere and circumnavigate each pole.

During winter the northern hemisphere usually has three or four long waves. The latitudinal thermal gradient is maximized during winter when the polar areas receive little to no insolation and the low latitudes still receive about as much energy as it receives in summer. Rossby waves must have some amplification to redistribute this energy imbalance latitudinally in an attempt to equalize the differences. In summer the number of long waves increases to as many as six, but the amplitudes decrease because the latitudinal thermal gradient decreases when the high latitudes warm up under long day lengths. In both winter and summer Rossby waves tend to be less meridional in the southern hemisphere than in the northern hemisphere. Instead, the near-continuous ocean in the southern hemisphere plays a greater role in the redistribution of energy than it does in the northern hemisphere.

Because these upper-air Rossby waves result from latitudinal thermal inequalities, the ridges and troughs are essential in the energy balance of a hemisphere. The trough-to-ridge sides of these waves allow warm, and often moist, low-latitude air masses to move poleward, and the ridge-to-trough sides of Rossby waves allow cold and often drier air to move equatorward. Without these long waves helping to redistribute energy across the midlatitudes, energy differences would continuously increase between the tropics and the poles. This would result in the poles becoming increasingly colder while the equator increasingly warmed, and soon life would be restricted to the very narrow zone where a thermal balance was achieved.

Vorticity

An important feature of upper-atmospheric Rossby wave flow in climatology is **vorticity**—the rotation, or spin, of any object. Although seemingly simplistic, vorticity may be viewed in a number of ways. The vorticity of an object such as a person, a particle in the atmosphere, or a storm system can occur either because the object itself is rotating or because it is situated on or in another object that is rotating (i.e., Earth), or both. **Relative vorticity** (usually denoted by ζ, the lowercase Greek letter zeta) is the spin that occurs because the object itself is turning. If you hold your pen vertically and twirl it between your fingers, you impart relative vorticity on the pen. If you hold your pen stationary but instead *you* begin rotating, your pen is still rotating but for a different reason.

Particles in the atmosphere are always rotating because they possess **planetary vorticity** (f)—the spin they acquire because of the rotation of Earth. The amount of this vorticity is proportional to the Coriolis effect, which as we have seen is itself proportional to latitude and speed of motion. Specifically, in the expression of the Coriolis effect from Chapter 3,

$$CE = 2\,\Omega\,(\sin\,\varphi)\nu,$$

the term $2\Omega(\sin\varphi)$, where Ω is the angular velocity of Earth and φ represents latitude (with negative values for southern hemisphere latitudes), is sometimes referred to as the **Coriolis parameter**, but it really represents f.

Earth rotates on its axis in such a way that if we were to look down from space directly on the North Pole we would see counterclockwise rotation (from west to east). But when viewed from below the South Pole, Earth is simultaneously rotating clockwise. By convention, any rotation in the same direction as Earth's rotation in that hemisphere is termed **positive vorticity**. Thus, counterclockwise-spinning objects in the northern hemisphere and clockwise-spinning objects in the southern hemisphere (both of which include cyclones in their respective hemispheres) are said to exhibit positive or "cyclonic" vorticity. Thus, for residents of either hemisphere, "positive vorticity advection" suggests the approach of a storm. Objects spinning in a clockwise manner in the northern hemisphere (or counterclockwise in the southern hemisphere) are said to have **negative vorticity**, because this rotation opposes that of Earth itself. Anticyclones in either hemisphere have negative vorticity.

In the northern hemisphere air flowing around ridges also has negative vorticity, because it is moving clockwise around the ridge, while air moving around troughs has positive vorticity. In the southern hemisphere ridges are also associated with negative vorticity (as the air moves counterclockwise through the ridge) and troughs are linked to positive (clockwise) vorticity.

Usually, vorticity in the middle atmosphere (i.e., the 500-mb geopotential height level) is considered to be most important for analysis. This is because if positive (or negative) vorticity is occurring halfway up the mass of the atmosphere, that same spin is likely to be propagated upward and downward to encompass most of the mass of the atmosphere over that location. If that vorticity is positive, the likelihood for storminess is enhanced. If it is negative, storm development is less likely.

The shearing motions that result from variations in wind speed across a horizontal surface are described as **transverse wind shear**, a force that is responsible for much of the ζ in air and can lead to positive or negative vorticity (**Figure 7.9**). To understand how transverse wind shear can cause ζ to exist in the Rossby waves, imagine a fast-meandering core of air, represented as a place where the pressure gradient is steeper than elsewhere in an upper-level weather map (**Figure 7.10a**). As this jet stream moves across a horizontal

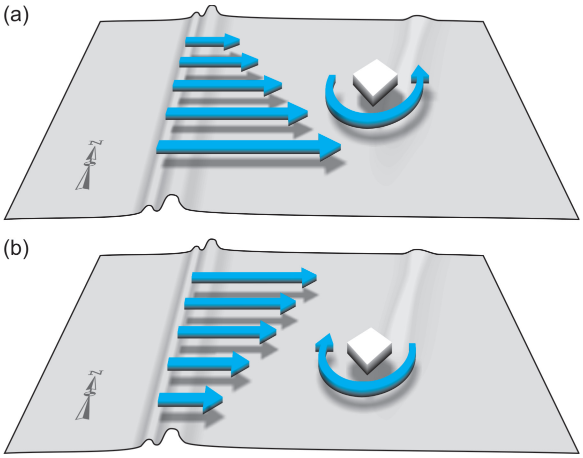

Figure 7.9 Transverse wind shear that produces (a) counterclockwise rotation, or positive ζ, and (b) clockwise rotation, or negative ζ.

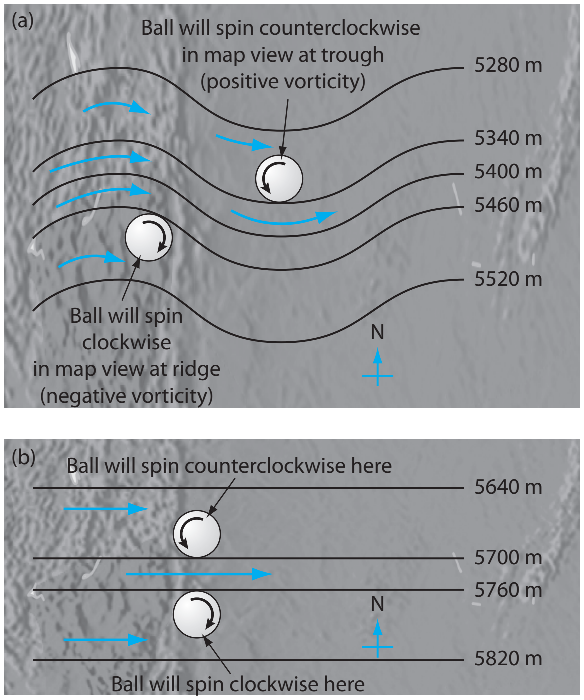

Figure 7.10 An example of the impact of transverse wind shear on relative vorticity (a) in a Rossby wave and (b) in a zonal flow.

surface in the midtroposphere in the area approaching a ridge, the air is pushed poleward. If we place a giant ball on the inside of the curve in the jet stream at the ridge, it would rotate against the spin of Earth's surface—in a clockwise direction in the northern hemisphere. This is because transverse shear is created from the north side of the ball in the northern hemisphere (where air is moving quickly) to the south side of the ball (where air is moving slower). Thus, the ridge area is characterized by negative ζ (regardless of hemisphere). Now imagine a position downstream from the ridge. There the jet stream meanders such that the belt dips equatorward. If we place a giant ball on the inside of this trough at the jet, the ball would spin in a counterclockwise manner in the northern hemisphere. This causes positive ζ (regardless of hemisphere) as a result of transverse wind shear.

If we track a parcel of air moving from a ridge to an adjacent downstream trough, we find that ζ increases as the air parcel moves downstream, regardless of hemisphere. Air near the ridge axis experiences the strongest negative ζ (in either hemisphere) because the parcel spins maximally in a clockwise (or counterclockwise in the southern hemisphere) manner at that location. Air near the trough axis experiences the maximum amount of positive ζ because a parcel spins maximally in a counterclockwise (or clockwise in the southern hemisphere) manner at that location. Between the ridge and the trough, air parcels show a ζ increase from negative to positive values in either hemisphere.

As the parcel continues toward the next downstream ridge, ζ begins to decrease as the air moves from a positive ζ region (the trough axis) toward a negative ζ region (the ridge axis) in either hemisphere. Again, this ζ change is caused by transverse wind shear. At the midway point between the trough axis and the ridge axis, the parcel of air experiences zero ζ, after which the rotation reverses.

Because ζ is a function of airflow curvature, strongly meridional patterns show the greatest ζ changes between ridges and troughs. Zonal patterns have only small ζ changes between the ridge and trough axes if the pressure gradient is constant. However, changes in the latitudinal pressure gradient cause transverse wind shear even if the flow is zonal (Figure 7.10b). Rossby wave patterns provide both favorable and unfavorable zones for cyclogenesis, which requires counterclockwise rotation (in the northern hemisphere) or positive ζ (in either hemisphere). Deamplified or zonal flow provides neither strong support for nor resistance to cyclogenesis, except when differing wind speeds across space cause transverse shear.

Changes in ζ as air moves also impact the development and life cycles of weather systems. If the air acquires positive ζ, the formation of low pressure and cyclogenesis is encouraged. Surface anticyclones are supported by negative ζ in either hemisphere. This occurs at 500-mb ridges. Surface cyclones, such as those manifested as midlatitude storm systems, are supported by 500-mb troughs. The greater the amount of positive and negative ζ present aloft, the greater the spin of air through the column of air between the middle troposphere and the surface. This is associated with the strength of the individual surface systems and their resulting atmospheric features.

The characteristics of ζ are so closely related to surface system characteristics that most weather forecasters examine a map of 500-mb geopotential heights and ζ to determine initial forecasting scenarios. In the case of midlatitude storminess, low geopotential heights combined with positive vorticity advection (in either hemisphere) indicate the position of the core of a cyclone in the secondary circulation. Numerical models project changes in ζ and positions of the ζ maxima over time. If the ζ in the region shows signs of maintaining itself or strengthening in the computer models, then the surface cyclone is likely to either maintain its strength or increase in strength accordingly. If the midtropospheric ζ in a region shows signs of weakening, then so will the surface system, ultimately to the point of dissipation. The future trajectory of the surface system may be forecasted by using computer-projected 500-mb vorticity characteristics. Similar analyses are conducted at climatological time scales for forecasting potential impacts of long-term changes in the general circulation on precipitation and storminess across space. Analogous modeling endeavors have estimated climatic conditions in the distant past.

The climate of a region is determined by the frequency and distribution of weather systems (and, therefore, vorticity) passing over it. However, ζ is only one component of the total vorticity. Even in the total absence of transverse wind shear (hypothetically a perfectly zonal pattern with no changes in pressure gradients), f is present in the moving atmosphere, except at the equator where the Coriolis effect is zero. **Absolute vorticity** (denoted by N, the uppercase Greek letter nu) is the sum of the relative and planetary vorticity. Expressed quantitatively,

$$\mathrm{N} = \zeta + f$$

The most important feature of N is that it is conserved—that is, it remains constant—in the atmosphere. This property has important ramifications for planetary-scale atmospheric flow.

Constant Absolute Vorticity Trajectory

As we have seen, cyclones have converging air motions at the surface but offsetting diverging air motions aloft (regardless of the hemisphere). Anticyclones have diverging air motions at the surface but have offsetting converging air motions aloft. The upper-air level where atmospheric motions oppose those near the surface maximally is typically located just beneath the tropopause. Somewhere between the two zones is the **level of nondivergence**, where neither convergence nor divergence occurs. Although the level of nondivergence may vary, the 500-mb geopotential height surface is often a good approximation of the level of nondivergence, because it is located where about half of the mass of the atmosphere is above it and half is below it.

Carl Rossby showed that at the level of nondivergence air motion follows a **constant absolute vorticity trajectory**. A parcel of moving air simply follows a path that conserves N over time. Imagine a parcel moving laterally in the midlatitude westerlies along the level of nondivergence. If it begins to move higher in latitude at all—say from southwest to northeast in the northern hemisphere (or northwest to southeast in the southern hemisphere)—Coriolis deflection is increasing (Figure 7.11). At the same time f is increasing, to conserve N the amount of ζ must decrease as it moves poleward. This causes it to acquire negative ζ (clockwise flow in the northern hemisphere), thereby creating

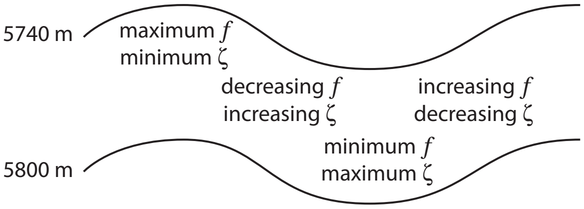

Figure 7.11 Relative and planetary vorticity features associated with midtropospheric Rossby waves in the northern hemisphere.

a ridge (in either hemisphere). When the parcel reaches the area of maximum negative vorticity (and maximum f)—the ridge axis—its clockwise (in the northern hemisphere) flow begins to take it toward the southeast (or northeast in the southern hemisphere). N is conserved.

As the parcel continues to move southeastward (northeastward in the southern hemisphere), its f decreases; so ζ must increase. This increasing ζ manifests itself as counterclockwise flow in the northern hemisphere (clockwise in the southern), and continued counterclockwise flow (clockwise in the southern hemisphere) eventually takes the parcel to a point at which it is as far equatorward as it will go. That point is the one at which f is at a minimum and ζ is at its maximum—the trough axis. Again, N is conserved. Continued motion toward the next downstream ridge once again sees increasingly positive f and increasingly negative ζ in either hemisphere.

Changes in ζ and Coriolis deflection cause adjustment in motion throughout the entire wave train of air around the hemisphere. For instance, as air moves into a ridge axis, Coriolis deflection increases as the parcel moves into higher latitudes. Even though there is an offset of ζ, the Coriolis deflection to the right in the northern hemisphere (left in the southern) causes the air parcel to turn equatorward on exiting the axis. The parcel moves toward lower latitudes along a northwest to southeast trajectory (or southwest to northeast trajectory in the southern hemisphere). At the trough axis f is at its weakest point, while the amount of ζ is maximized. This ζ overshadows the Coriolis deflection such that the parcel now turns to the left in the northern hemisphere (right in the southern hemisphere), causing the air to move along a southwest to northeast trajectory (northwest to southeast in the southern hemisphere). The result is that a ridge of a particular amplitude is associated with downstream adjustments that affect the development of a trough. The process continues with N being conserved as a function of ever-changing Coriolis deflection and ζ values. A series of long waves circumnavigating the entire hemisphere results.

Because N is conserved, any change in airflow at any location along the wave pattern is "felt" by all points upstream and downstream. For instance, if air were to speed up over a trough-to-ridge side of the Rossby wave in a meridional flow pattern, the Coriolis effect would increase because it is proportional to wind speed (and because f has increased). This increased pull to the right (in the northern hemisphere) would be instantly offset by a decrease in ζ to conserve N. Such a situation would amplify the ridge. The flow would then readjust (again) because increasingly meridional motion would cause more extreme decreases in f with equatorward flow, initiating rapid increases in ζ to conserve N. This would cause the downstream trough to deepen, and meridional flow would be further intensified. Adjustments in this area would trigger further readjustments in the next downstream ridge and so on, leading to an increasingly meridional pattern over time. Any adjustment of airflow in one location is propagated downstream until the entire longwave pattern fully readjusts to the new stimulus.

At some point a critical maximum is reached when sufficient energy has been redistributed meridionally. At that time the pattern readjusts to a more zonal pattern. Such a situation often creates closed migratory low- and high-pressure areas, which represent airflow regions very similar to cutoff meanders in streams.

When we view Rossby wave patterns over time, we see preferred geographical areas where flow adjustments occur. Pairs of locations where ridging and troughing or opposite pressure anomalies frequently occur simultaneously are termed **action centers**. Any adjustment in the pressure pattern in an action center triggers an offsetting adjustment at the other action center. Likewise, changes over time occur in preferred locations of ridge–trough pairs along the longwave pattern.

The correlated atmospheric flow patterns resulting from a "see-saw" pattern of pressure and ridging/troughing at action centers are termed **teleconnections**. Because teleconnections are associated with adjustments in midtropospheric patterns, they are responsible for driving climatic variations at the surface for many regions.

Flow Over Mountainous Terrain

Changes in flow pattern characteristics can occur for reasons other than conservation of N. For instance, in mountainous areas the atmosphere constricts vertically. This occurs as the space between the surface and the tropopause (h) decreases as surface elevation increases (**Figure 7.12**). When a moving fluid constricts to a smaller vertical

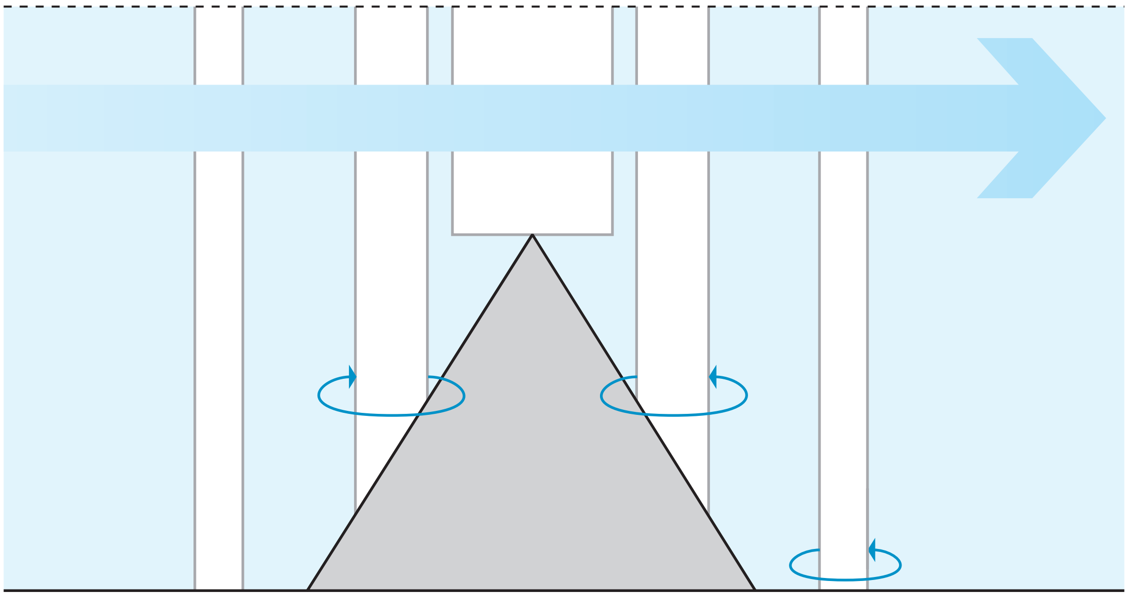

Figure 7.12 Prevailing airflow conserving potential vorticity encountering a mountain range.

space, a speed change occurs because, like N, **potential vorticity** is conserved. Potential vorticity can be expressed as

$$\mathrm{PV} = \frac{\zeta + f}{\Delta h} \quad \text{or} \quad \mathrm{PV} = \frac{\mathrm{N}}{\Delta h}$$

Consider the case of air moving from west to east in the midlatitude westerlies across western North America, when it encounters the Rocky Mountains. As air moves upslope Δh decreases, so N must also decrease. But if the air is flowing from west to east in the midlatitude westerlies, f is unchanging because f is a function of latitude, which does not change in west-to-east flow. For potential vorticity to be conserved, ζ must decrease as the air moves upslope. As we have seen, decreasing ζ is associated with increasingly anticyclonic (clockwise in the northern hemisphere) flow. This encourages a ridge to form immediately west of a high mountain range that is oriented perpendicular to the direction of the flow.

By contrast, as air flows downslope on the *leeward* side of the mountains, Δh increases, so N must also increase (Figure 7.12). But again, f is constant for westerly flow, so for potential vorticity to be conserved, ζ must increase. Increasing ζ is associated with cyclonic (counterclockwise in the northern hemisphere) flow. This encourages a trough to form downstream from the peaks in the midlatitudes in each hemisphere. Cyclogenesis is common leeward of major midlatitude mountain ranges.

Baroclinicity

Another situation that leads to changes in Rossby wave flow involves **baroclinicity**—the intermixing of steep thermal gradients across small regions. Such gradients often occur at coastlines, where there is a differential heating of land masses versus ocean surfaces at a given line of latitude. The most pronounced thermal gradients occur on the leeward side of midlatitude continents during winter as the continents chill substantially while the ocean temperatures remain relatively mild. This effect is reinforced by the warm ocean currents that dominate the western sides of ocean basins (eastern sides of continents) as low-latitude waters move poleward in the oceanic **gyres**. During winter the thermal gradient across the eastern coasts of continents is maximized.

Any region (at the surface or aloft) where air is being advected into a region with a drastically different temperature is termed a **baroclinic zone**. Baroclinic zones are regions of active and sometimes severe weather in the midlatitudes. Geographical areas where baroclinic zones are preferred have frequently stormy climates. Low-level baroclinicity generally causes wind speed to strengthen with elevation as the heights of constant upper-air pressure surfaces attempt to even out over a small region. These changes manifest themselves through the longwave pattern. In general, if a baroclinic zone occurs in association with an upper-level trough, the development of surface cyclones is strongly encouraged. Baroclinic zones are apparent on a map where isotherms are packed closely together, especially when they are nearly perpendicular to the Rossby wave flow. The latter indicates that efficient thermal advection is occurring. In **Figure 7.13**, notice how the 500-mb baroclinic zone focused over Utah and Idaho is associated with the longwave trough. As air moves through that trough, it acquires counterclockwise rotation (i.e., positive vorticity) and supports cyclonic development.

Rossby Wave Divergence and Convergence

Surface cyclones require diverging air aloft to sustain surface convergence. Such air is found on the trough-to-ridge side of the Rossby waves (**Figure 7.14**). To see why, imagine air moving around the trough aloft (in the northern hemisphere). In such a case it must turn to the left to remain parallel

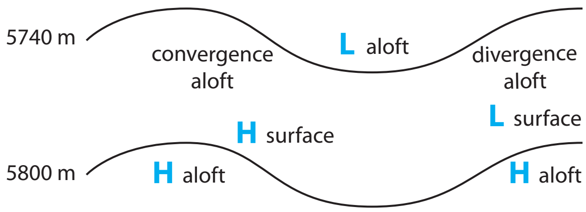

Figure 7.14 Areas of favorable surface cyclone and anticyclone formation relative to the Rossby wave pattern.

to the **isohypses**—lines of constant geopotential height. Flow in the upper-level maps generally parallels the isohypses because *geostrophic balance*—a balance between the pressure gradient force and the Coriolis effect—occurs aloft. To turn against the direction induced by the Coriolis effect, the speed must be slow enough to reduce the Coriolis effect sufficiently.

On the other hand, as air moves through a ridge (in the northern hemisphere) it turns to the right to remain parallel to the isohypses. Because this is the direction that the Coriolis effect deflects the air, the speed increases coincident with the strengthening of the Coriolis effect. Note the Coriolis effect weakens at the trough and strengthens at the ridge (in either hemisphere) because latitude is relatively low at the trough and higher at the ridge.

Thus, air diverges on the trough-to-ridge side of the Rossby wave aloft, moving relatively slowly around the trough and then relatively fast around the downstream ridge (assuming that the isohypse gradient is the same for both). This divergence on the trough-to-ridge side of the upper-level Rossby wave induces rising motion to replace the air that has "spread out" aloft. Air cannot approach laterally because it must remain parallel to the isohypses to ensure the balance of forces suggested by geostrophic flow. The rising motion is supportive of surface low pressure—a cyclone. Therefore, surface cyclones are found on the trough-to-ridge side of the wave in either hemisphere (Figure 7.14).

By contrast, the ridge-to-trough side of the wave is associated with convergence aloft (in either hemisphere) as air slows down and "piles up." This abundance of air aloft induces a sinking motion on the ridge-to-trough side of the wave and surface high pressure. Thus, surface anticyclones are located beneath the ridge-to-trough side of the Rossby wave in either hemisphere.

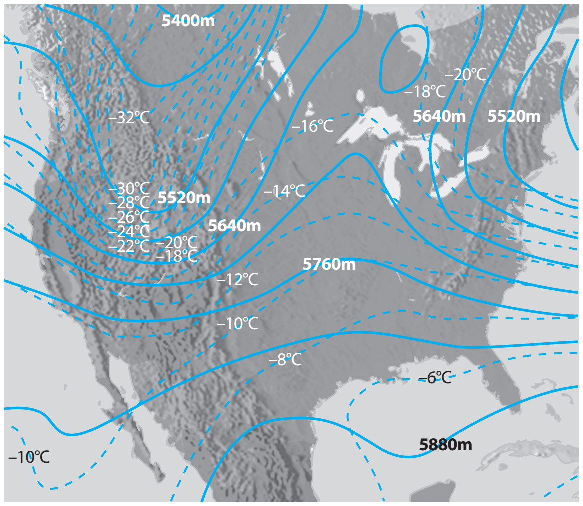

Figure 7.13 A 500-mb weather map for April 18, 2006. Notice how the isotherms (dashed lines) are packed closely together in the Mountain West of North America on this day, coinciding with the closely packed isohypses (solid lines), creating a strong baroclinic zone on that day. By contrast, few isotherms appear in the southern United States where lighter 500-mb winds occur. *Data from*: The Ohio State University WWW Weather Center, courtesy of the Department of Geography/Atmospheric Sciences Program.

Rossby Wave Diffluence and Confluence

Additionally, areas of diffluence and confluence in the upper-level Rossby waves can create rising and sinking motion. **Diffluence** refers to the horizontal spreading of air streams at a given height, and **confluence** involves areas where air is converging horizontally (**Figure 7.15**). Upper-level diffluence supports rising air motions by developing an upper-air vacuum as air parcels spread away from each other. Air rushes into the vacuum from below, which initiates and maintains surface low-pressure centers.

Confluence consists of air streams that converge horizontally toward each other. At upper levels the "pile up" of air associated with such motions causes air to slow down. Gravity prevents this "extra" air from moving out to space, so the only place the air can move is downward. Thus, sinking motions result. These motions initiate and support the development of surface anticyclones.

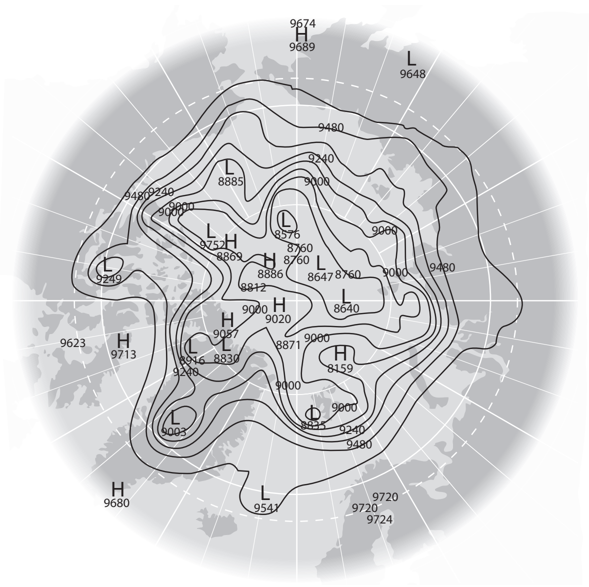

Figure 7.16 The northern hemisphere's circumpolar vortex at the 300-mb level on May 21, 2006. The locations where the isohypses are packed closely together represent the extent of the northern hemisphere's circumpolar vortex. *Data from*: National Weather Service/NOAA.

Polar Front Jet Stream

Embedded within the upper-level midlatitude westerlies are cores of extremely fast airflow. These places are indicated on upper-level maps by isohypses that become very closely spaced (but, of course, isohypses can never intersect each other). If we examine a map with the north (or south) pole in the center, as in **Figure 7.16**, an enclosed area of closely spaced isohypses would be apparent some distance from the pole over the midlatitudes. Airflow through these closely spaced isohypses is continuous around the hemisphere. The enclosed region represents the **circumpolar vortex**, an area of strong winds aloft that encircles the surface polar high. The *polar front jet stream*, which marks the periphery of the circumpolar vortex, is a core of fast-moving air that meanders through the upper troposphere. There is one polar front jet in the northern hemisphere and another in the southern hemisphere. The strongest wind speeds occur just beneath the tropopause in the 300-mb to 200-mb height field area where the geopotential height gradient is strongest and friction is minimal.

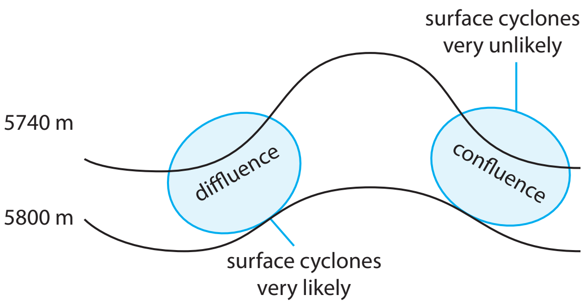

Figure 7.15 Areas of favorable surface cyclone and anticyclone formation relative to the Rossby wave pattern: diffluence and confluence.

The region of greatest wind speed within the polar front jet stream—the **jet streak**—occurs over the primary baroclinic zones, which exist at the boundary between cold air of polar origin and equatorially derived air. The speed of the polar front jet stream is directly proportional to the latitudinal thermal gradient, with steeper thermal gradients and baroclinic zones inducing steeper geopotential height gradients and a faster jet. Baroclinicity supports and maintains jet stream characteristics.

Air of tropical and polar origin meets in the midlatitudes to produce active baroclinic zones. The tropical air is usually also humid, because a large geographical area of the tropical Earth is over oceans and because warm air has high *saturation vapor pressure*. The polar air is often also dry, because relatively more of the polar Earth is either covered by land or sea ice and because the saturation vapor pressures decrease with decreasing temperatures. The boundary between the tropical and polar air pools produces a distinct thermal–moisture boundary through the vertical profile of the atmosphere. Air temperatures change significantly over very small areas near the boundary. The location of this thermal gradient at a given time determines the location and shape of the circumpolar vortex and the location of the polar front jet stream in the zone between the subpolar lows and the subtropical highs.

The height of geopotential pressure surfaces is directly related to the thermal properties of air, in fulfillment of the *equation of state*, also known as the *ideal gas law*, for dry air:

$$P = \rho R_d T$$

where P represents pressure (in *Pascals*), ρ represents density, R_d is the dry gas constant, and T is in the *Kelvin temperature scale*. If we were to examine the 500-mb geopotential height field (an unchanging pressure value), we would notice that the density of that pressure surface decreases over areas of higher temperature and increases over areas of lower temperature, in fulfillment of the equation of state. Density decreases when air rises, so warmer areas must have higher constant pressure surfaces than colder areas (Figure 7.4). This is a direct result of thermal expansion (warming conditions) and contraction (cooling conditions) of a fluid, as described by the equation of state.

A sharp thermal gradient occurs between warm and cold air at the surface with the boundary extending through a vertical profile of the atmosphere. In the vicinity of the 300-mb geopotential height layer and extending to about the 200-mb layer (just beneath the tropopause), geopotential heights slope sharply across the thermal boundary. Remember that air in the upper atmosphere flows down the geopotential thermal gradient from equatorward to poleward latitudes. The pressure gradient force is always pulling air toward the poles as the geopotential height fields are lower in height (thermally contracted) than over lower latitudes (thermally expanded). This creates a very steep geopotential height slope above the thermal boundary between warm and cold air.

As air begins to flow across this gradient, it accelerates. This increase in slope provides energy to the jet stream. However, air does not simply flow over the sloping heights toward the pole. As we already know, Coriolis deflection pushes air to the right in the northern hemisphere, allowing for geostrophic balance to be reached with the pressure gradient force, which is pulling air toward the pole. This results in a rapidly moving stream of air flowing generally from west to east above the warm–cold air boundary. The same situation (heights sloping poleward) occurs in the southern hemisphere, which also results in a west-to-east-flowing jet stream as air is pulled to the left by Coriolis deflection.

The polar front jet stream's location varies as the flow of air travels along the area of constant N. The jet stream as identified on upper-air weather maps can be used to identify ridges and troughs, and corresponding Rossby waves, in the upper troposphere. The jet is also important in governing surface weather, because it represents the boundary between cold and warm air and it tends to govern the development and movement of midlatitude storm systems. Analysis of the polar front jet stream and prediction of its future motions are essential for short-term and long-term atmospheric prediction and analysis.

The thermal boundary associated with the polar front jet stream is generally much more prominent in winter because the latitudinal thermal gradient is much greater during that season than during the summer. Wind speeds typically reach 160 km hr^{-1} (100 mi hr^{-1}) in winter but drop to about 80 km hr^{-1} (50 mi hr^{-1}) in summer.

Mean Patterns of Rossby Wave Flow

Climatological analysis of the polar front jet stream is often done by examining average patterns, even though variability and extremes are important components of climatology. The jet stream's position varies substantially from day to day, but analysis of the position and intensity of the jet over long time scales shows distinct monthly and seasonal patterns.

As explained above, the sharp surface thermal gradients associated with coastal locations produce baroclinic regions that generally correspond to the location of the polar front jet. The constraint of upper airflow by conservation of absolute vorticity results in rather distinct atmospheric teleconnections. The polar front jet is located maximally equatorward during the coldest part of the year. This results from the expansion of the polar anticyclone and subtropical lows and contraction of the subtropical anticyclones during winter. In summer the semipermanent pressure cells reverse from their winter tendencies. This causes the polar jet to retreat maximally poleward.

The presence of topographical barriers and baroclinic zones causes persistence in the long-term jet pattern. For instance, air motion over North America is typically characterized by a ridge over the western mountain cordillera and a trough over the east. This pattern is associated with the conservation of potential vorticity as the height of the air parcel decreases as it moves upslope and then decreases as it moves downslope. The amplitude of the jet changes in association with vorticity and thermal characteristics as described above. A reverse flow—trough in the west and ridge in the east—does occur, but it is rather infrequent. Recent studies show that once the reverse pattern establishes itself, it can be quite persistent.

Climatological averages of middle-tropospheric patterns for the northern hemisphere reveal many of the principles discussed above. **Figure 7.17** shows that 500-mb geopotential heights are generally higher in summer than in winter over any given location and that the 500-mb geopotential height gradients are steeper in winter than in summer. The equatorial areas have very weak 500-mb geopotential height gradients year-round. In contrast, few major seasonal changes occur in the southern hemisphere. The dominant presence of water, resulting in a fairly stable latitudinal thermal gradient in the southern hemisphere, keeps the patterns fairly constant throughout the year.

Summary

General circulation exists to redistribute energy that arrives at Earth in greater quantities near the equator than near the poles. Because of the rotation of Earth, this circulation is characterized by a Hadley cell and a polar cell in both the northern and the southern hemispheres. Air in the Hadley cell moves upward near the equator at the Intertropical Convergence Zone, with a poleward motion to about 30°N and 30°S latitude; sinks at the subtropical anticyclones; and then moves equatorward again in the form of the northeast trade winds (northern hemisphere) and the southeast trade winds (southern hemisphere). The subtropical jet stream exists aloft at the poleward edge of the Hadley cell in each hemisphere as angular momentum is conserved, and it exerts significant influences on the subtropical atmosphere, many of which are poorly understood.

The polar cell is characterized by sinking air at the pole, equatorward surface motion to about 60°N or 60°S (deflected by the Coriolis effect to become the polar easterlies), rising air at the subpolar lows, and then poleward motion (deflected to become westerlies) aloft. Between the Hadley and polar cells in each hemisphere are surface westerly winds, with stronger westerlies aloft. Within the midlatitudes atmospheric flow aloft takes the form of Rossby waves, which pull air toward the pole (ridge) and toward the equator (trough) as it moves generally from west to east embedded in the westerlies.

Vorticity is an important feature of atmospheric motion. Air flowing in upper-level Rossby waves conserves its absolute vorticity, thereby ensuring that ridges and troughs remain confined to the middle latitudes. Because potential vorticity is conserved, air moving upslope in the midlatitude westerlies acquires anticyclonic (negative) vorticity, while that moving down the leeward slope acquires positive vorticity. The trough-to-ridge side of the wave and any areas of diffluence along the wave support surface cyclones and cyclogenesis. The ridge-to-trough side of the wave and any areas of confluence along the wave support surface anticyclones and anticyclogenesis.

The circumpolar vortex and in particular the polar front jet stream exist over the zone of the sharpest temperature contrasts and the strongest baroclinic zone. This feature is embedded within the Rossby waves and can exert a significant influence on the weather and climates beneath it.

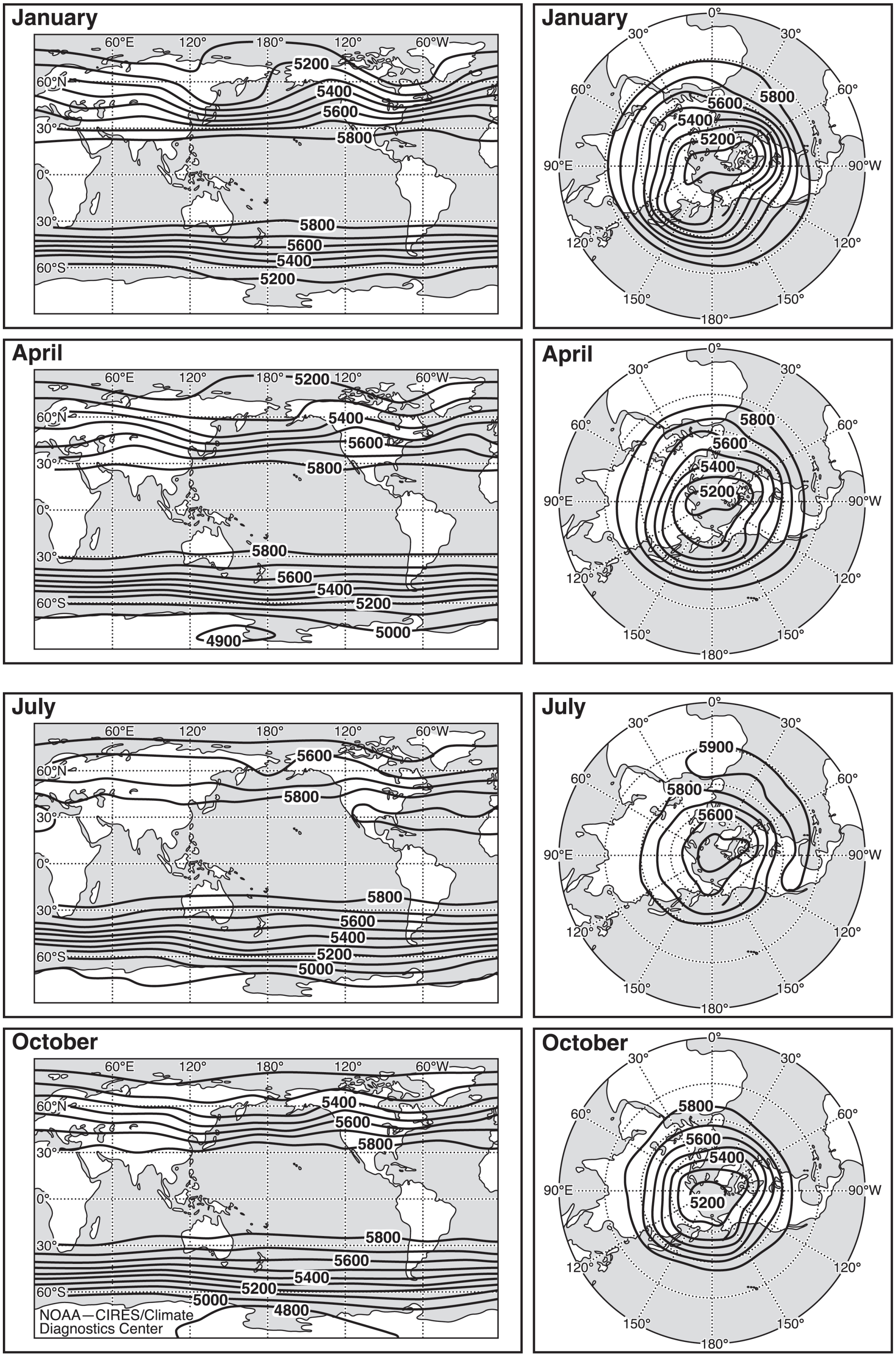

Figure 7.17 Mean 500-mb geopotential heights (1948–2004) in the northern hemisphere. *Data from*: NOAA/CIRES Climate Diagnostics Center and the University of Colorado at Boulder.

Key Terms

Absolute vorticity
Action center
Aerosols
Albedo
Aleutian low
Amplitude
Anticyclone
Baroclinic zone
Baroclinicity
Bermuda-Azores high
Circumpolar vortex
Confluence
Conservation of angular momentum
Constant absolute vorticity trajectory
Continentality
Convection
Coriolis effect
Coriolis parameter
Cyclogenesis
Cyclone
Density
Diffluence
Doldrums
Equation of state
Evaporation
General circulation
Geopotential height
Geostrophic balance
Gyre
Hadley cell
Hawaiian high
Horse latitudes
Icelandic low
Ideal gas law
Indian Ocean high
Insolation
Intertropical Convergence Zone (ITCZ)
Isohypse
Jet streak
Jet stream
Kelvin temperature scale
Kinetic energy
Latitude
Leeward
Level of nondivergence
Long wave
Longitude
Meridional flow
Midlatitude westerlies
Momentum
Negative vorticity
Newton's second law
North Pacific high
Northeast trade winds
Pascal
Planetary vorticity
Polar easterlies
Polar front jet stream
Polar cell
Polar high
Positive vorticity
Potential vorticity
Pressure gradient force
Relative vorticity
Ridge
Rossby wave
Sargasso Sea
Saturation vapor pressure
Second law of thermodynamics
Secondary circulation
Sensible heat
Siberian high
Solar declination
Solar noon
South Atlantic high
South Pacific high
Southeast trade winds
Specific heat
Stability
Subpolar low
Subtropical anticyclone
Subtropical jet stream
Teleconnection
Tibetan Low
Transverse wind shear
Tropic of Cancer
Tropic of Capricorn
Tropopause
Troposphere
Trough
Vorticity
Wavelength
Zonal flow

Terms in italics have appeared in at least one previous chapter.

Review Questions

1. Discuss the primary atmospheric circulation features present on a nonrotating planet with a continuous, uniform surface.
2. Describe the key pressure and wind features in the idealized circulation model.
3. Compare and contrast the idealized circulation model to reality. Why are they different?
4. Does the idealized global circulation model adequately describe upper-atmospheric motions? If not, why not?
5. Describe the waxing and waning of the semipermanent pressure cells through the course of the year.
6. Discuss the annual migration of the semipermanent pressure cells.
7. Identify and compare relative and absolute vorticity.
8. Discuss the role of positive and negative vorticity in association with ridges and troughs.
9. Describe the role of vorticity in the development of zonal and meridional Rossby wave patterns.
10. Why is absolute vorticity important? Of what is it a function?

11. Discuss the role of constant absolute vorticity trajectory relative to atmospheric motions aloft.
12. What are teleconnections and teleconnection action centers and why are they important to upper-atmospheric flow patterns?
13. Discuss the role of baroclinicity in the development of upper-atmospheric flow patterns.
14. What is the polar front jet stream, how does it form, and why is it an important feature of midlatitude upper airflow?

Questions for Thought

1. The text says that the Rossby waves in the southern hemisphere tend to be less meridional than those in the northern hemisphere. Why do you believe that this might be the case?
2. If the world were to warm up, what do you believe might happen to the various semipermanent pressure features around the world?
3. Assume an earth surface comprised solely of water. Describe midlatitude jet stream characteristics for this earth. Would troughs and ridges develop? Why or why not?

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