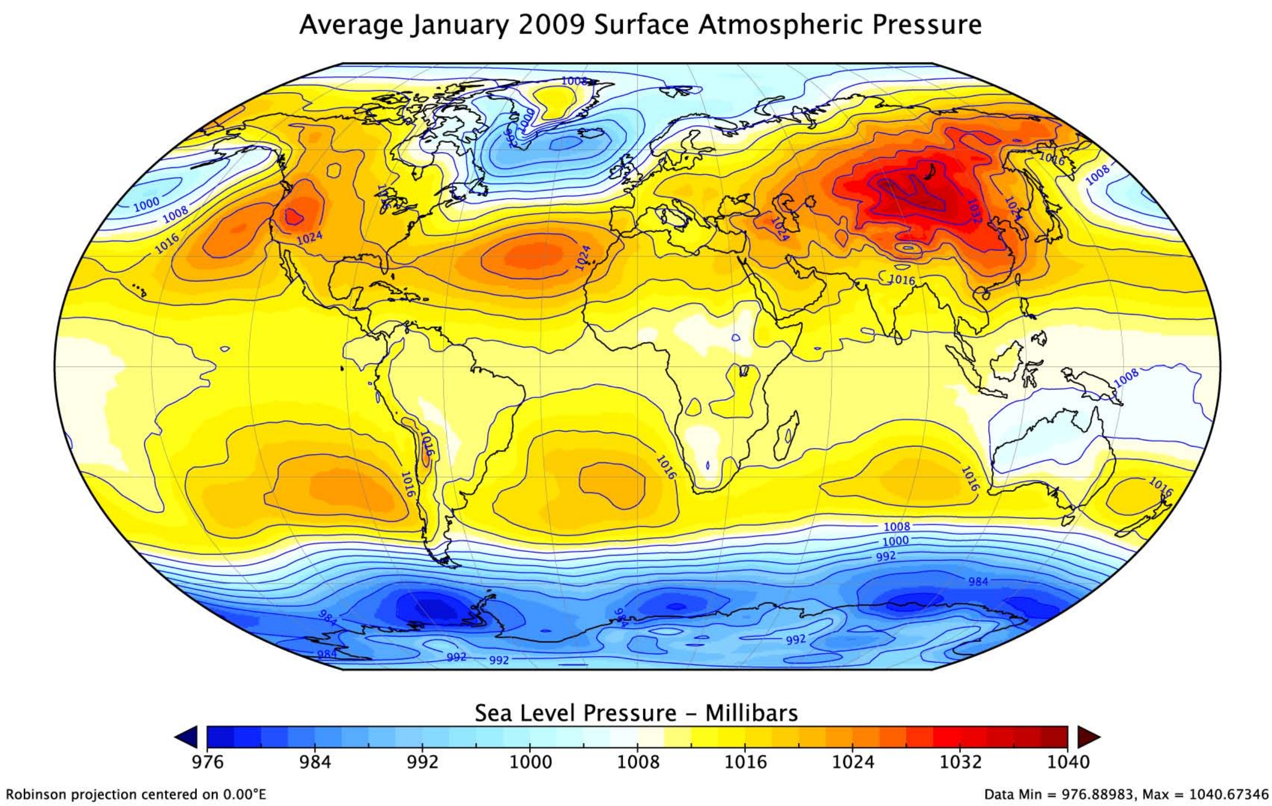
# Chapter 7: Atmospheric Pressure and Wind


## Michael Pidwirny


**Average January 2009 Surface Atmospheric Pressure**. The values displayed on this map have been adjusted to sea level. This map shows areas of high pressure over central Asia, western North America, and over the Pacific, Atlantic, and Indian Oceans at the subtropics. Areas of low pressure are found bordering Antarctica, south of Greenland in the Atlantic Ocean, and in the Pacific south of the Aleutian Islands. *(Data Source: NOAA - Earth System Research Laboratory)*

## STUDENT LEARNING OUTCOMES

### *After reading this chapter you should be able to:*

- Define the concept of atmospheric pressure and describe the instruments used to measure it.
- Describe the factors that cause atmospheric pressure to vary over time and across space.
- Define the concept of wind and describe how its wind speed and wind direction are measured.
- Outline how pressure gradient force, Coriolis effect, and frictional force influence winds near the Earth's surface and in the upper atmosphere.
- Describe the processes responsible for the formation of the following local to regional wind systems: land and sea breeze, mountain and valley breezes, and monsoons.
- Describe the mechanisms responsible for the formation of global scale surface and upper atmosphere circulation systems.
- Illustrate the average patterns of surface air pressure and wind at the global scale.

## ATMOSPHERIC PRESSURE

Located above the Earth's ocean and land surface is an extensive layer of gases, minute liquid droplets, and tiny solid particles that constitutes the atmosphere. All of these substances are forms of matter and therefore exhibit the properties of mass and weight. We can define **mass** simply as the amount of matter found in an object. Obviously, an equal volume of our planet's atmosphere and the materials found in ground beneath us (soil and rocks) do not have the same mass. As a rule of thumb, liquids have less mass than solids, and gases have less mass than liquids.

**Weight** can be defined as a measure of the gravitational force acting on an object. **Gravity** is a force that is found in all objects that have mass. The strength of this force is related to the size of an object's mass. Basically, the more mass an object has, the greater its gravitational force will be. The strength of this force is also influenced by distance, becoming less as one moves away from the center of the object. As a result, the strength of the gravitational attraction that occurs between any two objects in our Universe is determined by the mass of the objects and by the spatial distance that separates them.

The solid portion of the Earth has many times more mass than the atmosphere. Consequently, the atmosphere is pulled toward our planet's surface with great force. According to the previous discussion, the measurement of this force would be equal to the weight of the atmosphere's mass. This weight is routinely measured and is called **atmospheric pressure**. Near surface measurements of atmospheric pressure are made at meteorological stations all over the world several times a day. Some of these stations also regularly measure pressure vertically in the atmosphere using an electronic device attached to balloon called a **radiosonde** (this instrument also measures temperature and humidity). Measurements from the radiosonde are transmitted to a ground-based receiving station until the balloon expands so much that it pops. This normally occurs at a height of about 30 km (19 mi). Together, near surface and upper air measurements are used to help construct future forecasts of our planet's weather.

Atmospheric pressure varies both vertically and horizontally in our atmosphere. A number of factors are responsible for this fact. Most variations in atmospheric pressure are caused by changes in the amount of mass contained in the overlying atmosphere. Changes in atmospheric mass at the local or regional scale can occur with the movement of air because of atmospheric circulation (**Figure 7.1**). An increase in atmospheric pressure occurs when air collects in a particular location. Decreases in air pressure can arise when circulation removes air from an area.

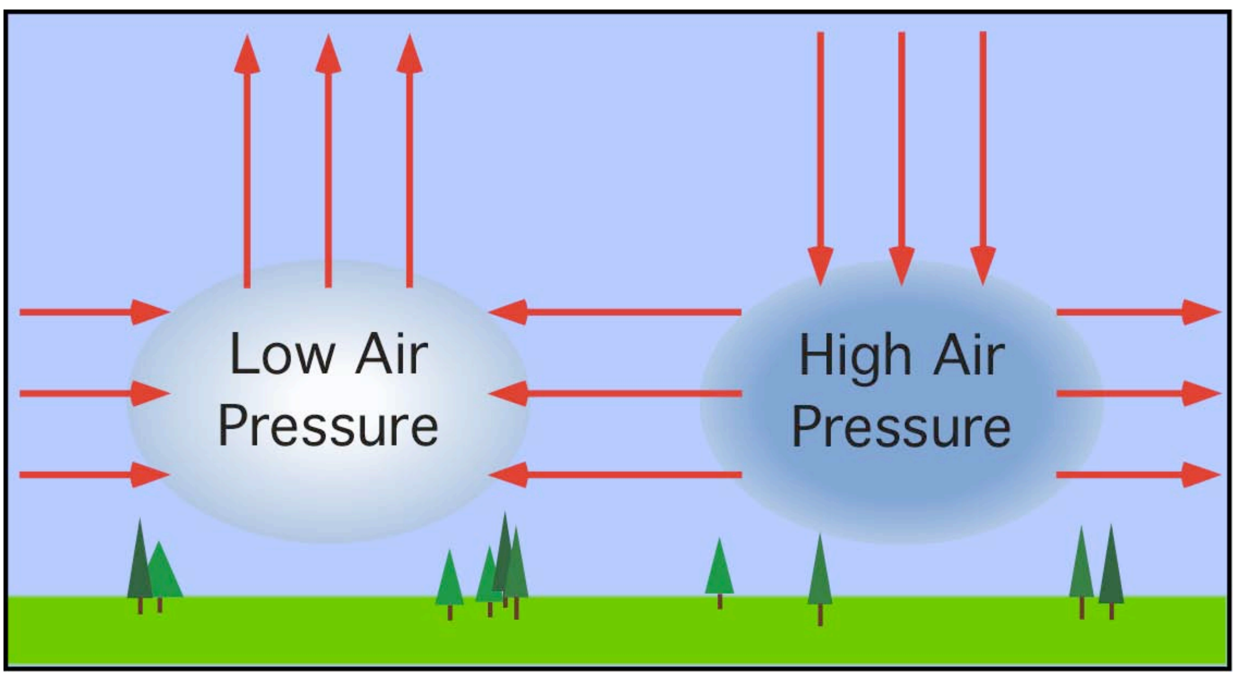



**FIGURE 7.1** Variations in atmospheric pressure can occur because of air circulation. In the illustration above, the movement of air out of a region in the atmosphere causes air pressure to drop. An increase in atmospheric pressure occurs when circulation patterns cause air to accumulate in a specific area of the atmosphere. *(Image Copyright: Michael Pidwirny)*

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Another factor that can alter atmospheric pressure is the concentration of moisture found in the air. The molecular weight of water vapor ($H_2O$) is actually less than the other two main constituents of the atmosphere, oxygen ($O_2$) and nitrogen ($N_2$). Consequently, increasing the concentration of water vapor found in a volume of air leads to a decrease in atmospheric pressure. Note that this fact is opposite to the popular belief that humid air is heavier than dry air.

## Pressure, Density, Volume, and Temperature

Changes in atmospheric pressure also can be caused by modifications in air temperature, volume, and density. The connection that exists between these variables and air pressure can be expressed mathematically. This mathematical relationship is known as the **Ideal Gas Law**. Two equations are commonly used to describe this law:

**Pressure x Volume = Constant x Temperature**

and

**Pressure = Density x Constant x Temperature**

From the equations shown above, we can define the following relationships between *pressure*, *density*, *temperature*, and *volume*. When *temperature* is held constant, increasing *volume* occupied by the gases of the atmosphere causes the *density* and *pressure* of the air to decrease, while decreasing the *volume* results in higher atmospheric *density* and *pressure* (**Figure 7.2**). If *volume* is kept constant, the *pressure* of a unit mass of the atmosphere is proportional to *temperature*. Consequently, an increase in *temperature* will cause *pressure* rise and a decrease in *temperature* will cause *pressure* to drop (**Figure 7.3**). Keeping *pressure* constant causes the *temperature* of a gas to be proportional to *volume*, and inversely proportional to *density*. Thus, increasing *temperature* of a unit mass of the atmosphere causes its *volume* to expand and its *density* to decrease, while a drop in *temperature* causes *volume* to decrease and *density* to rise (**Figure 7.4**).

## Surface Atmospheric Pressure

**Figure 7.5** shows the average change in atmospheric pressure with height above the ground surface. This figure indicates that as one moves from the Earth surface towards the edge of the atmosphere, the pressure of the atmosphere

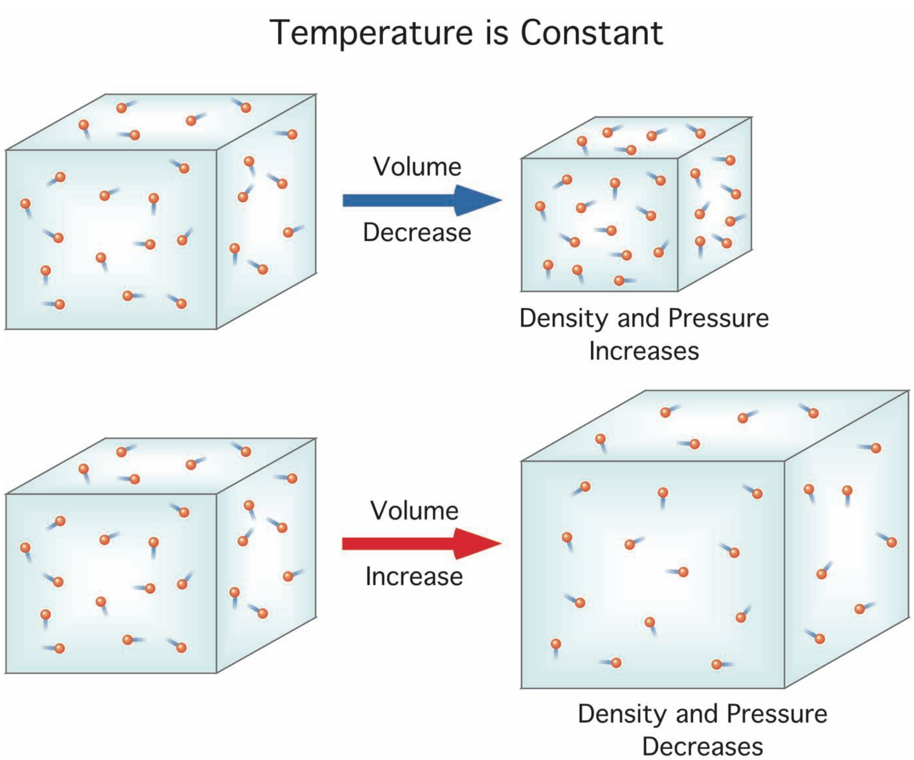


**FIGURE 7.2** Increasing or decreasing the volume occupied by a gas, while holding temperature constant, causes proportional changes in the density and pressure exerted by a gas. *(Image Copyright: Michael Pidwirny)*

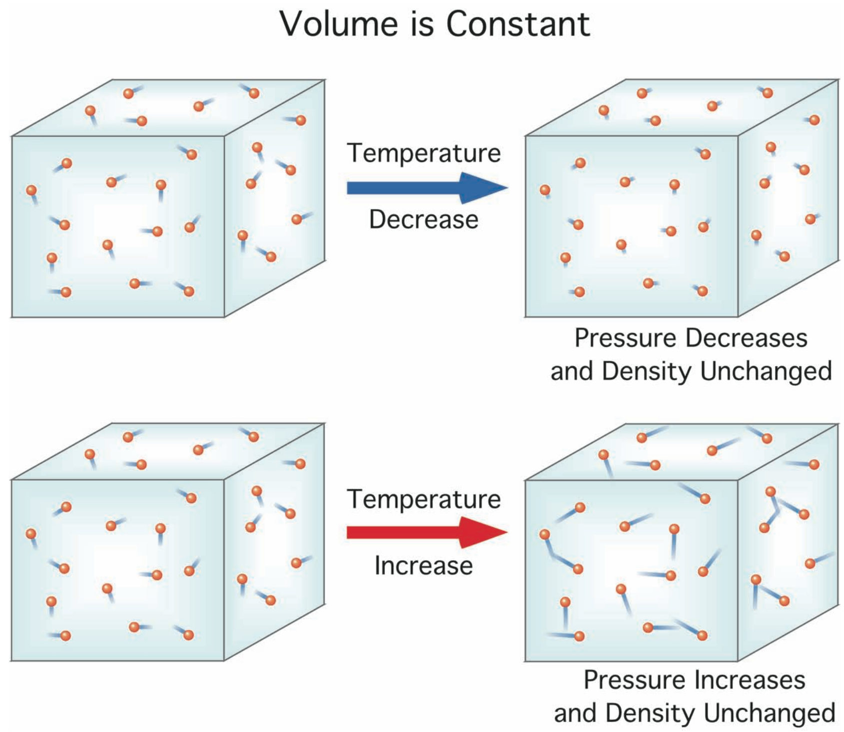


**FIGURE 7.3** Increasing or decreasing the temperature of a gas, while holding volume constant, causes proportional changes in the pressure exerted by a gas. The length of the trailing blue streaks indicates the relative speed of the moving gas molecules. *(Image Copyright: Michael Pidwirny)*

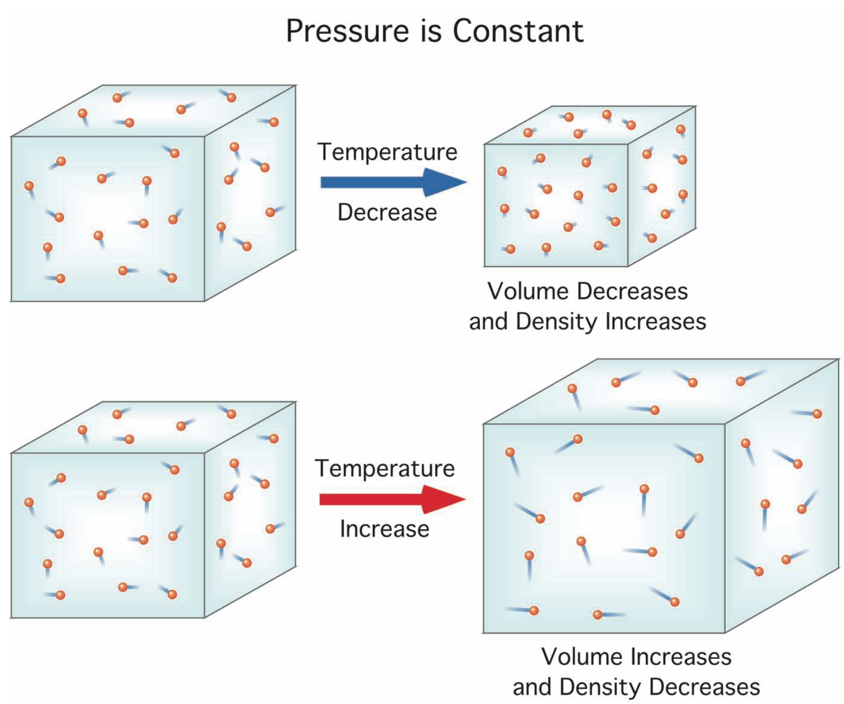


**FIGURE 7.4** Increasing or decreasing the temperature of a gas, while holding pressure constant, causes proportional changes in the density and volume of space occupied by a gas. The length of the trailing blue streaks indicates the relative speed of the moving gas molecules. *(Image Copyright: Michael Pidwirny)*

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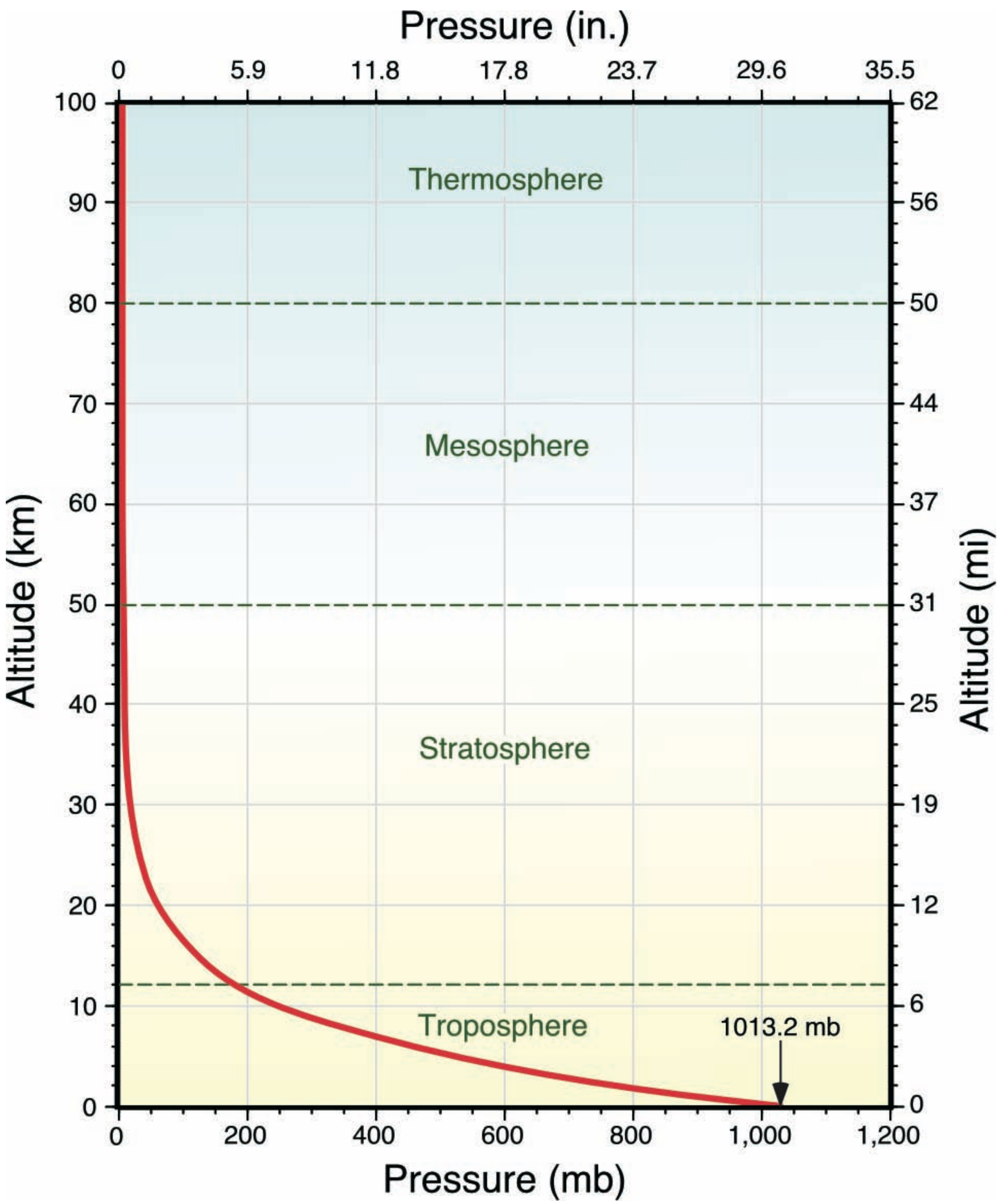


**FIGURE 7.5** Change in average atmospheric pressure with altitude. Note that the average atmospheric pressure at sea level is 1013.2 mb or 101.32 kPa. *(Image Copyright: Michael Pidwirny)*

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declines quickly with altitude. At the Earth's surface pressure averages about 1013.2 **millibars** (mb) or 29.92 inches (in.) of mercury. Both of these units are commonly used in the United States for quantifying atmospheric pressure. Physicists favor a metric unit known as a **Pascal** (Pa) for measuring pressure. One millibar is equal to 100 pascals. Meteorologists in countries that employ a metric system of measurements (like, Canada) often use a multiple of the pascal unit known as a **kiloPascal** (kPa). The prefix kilo means 1000. Consequently, one kiloPascal is the same as 1000 Pa or 10 mb. Another unit of force sometimes used by scientists to measure atmospheric pressure is the Newton (N). One millibar equals 100 Newtons per square meter ($N/m^2$ or $N\ m^{-2}$).

Variations in air pressure near the Earth's surface normally fall in a range from 970 to 1050 mb (**Figure 7.6**). Measurements in the upper half of this range (1010 to 1050 mb) are normally associated with clear skies, a dry atmosphere, and fair or cold weather. Pressure readings that are in the lower half of this range (970 to 1010 mb) are often related to cloudy skies, a moist atmosphere, and stormy weather. Air pressure values near the Earth's surface can occasionally fall outside the normal range just mentioned. For example, some tropical storms known as hurricanes can create localized areas of extremely low pressure. Hurricane Tip, which formed in the western Pacific Ocean during October 1979, was estimated to have a central pressure reading of 870 mb. This reading stands as the lowest pressure every measured at the surface of our planet. The highest recorded surface pressure occurred in Agata, Siberia during December 1968 with a reading of 1084 mb. This event was caused by an extremely cold and dry air mass that dominated Siberia on this date.

## Measuring Air Pressure

Any instrument that measures air pressure is called a **barometer**. The first measurement of atmospheric pressure began with a simple experiment performed by Evangelista

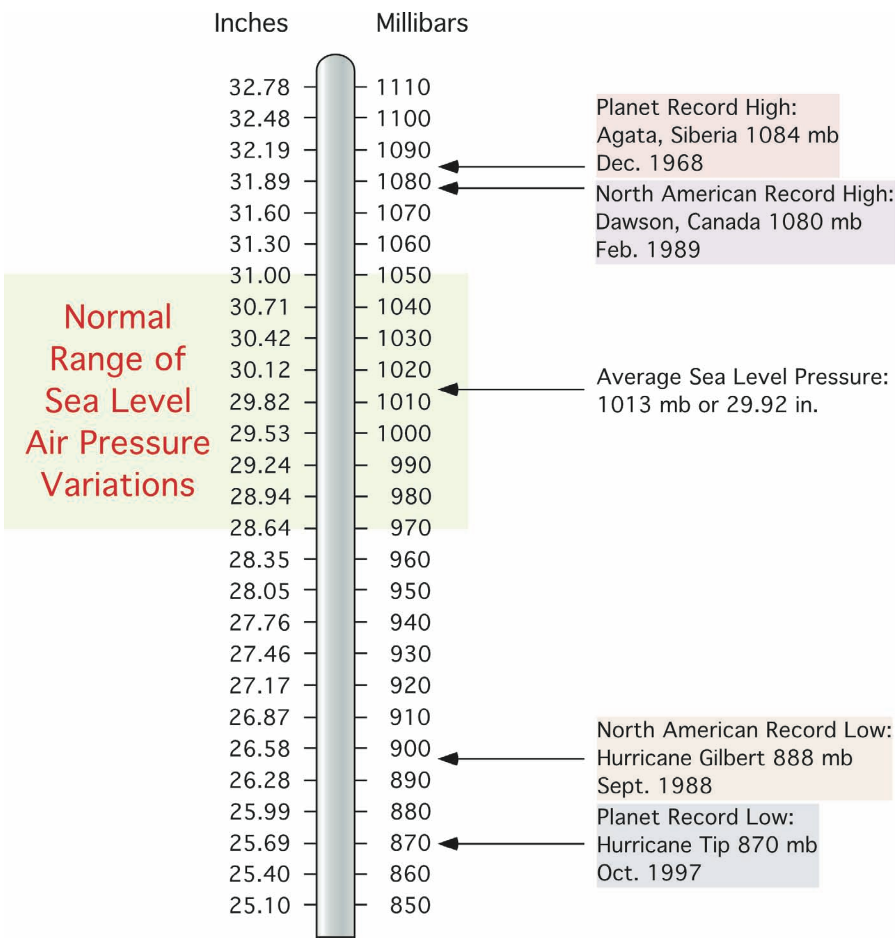


**FIGURE 7.6** Sea level variations and extremes in atmospheric pressure in inches of mercury and millibars. *(Image Copyright: Michael Pidwirny)*

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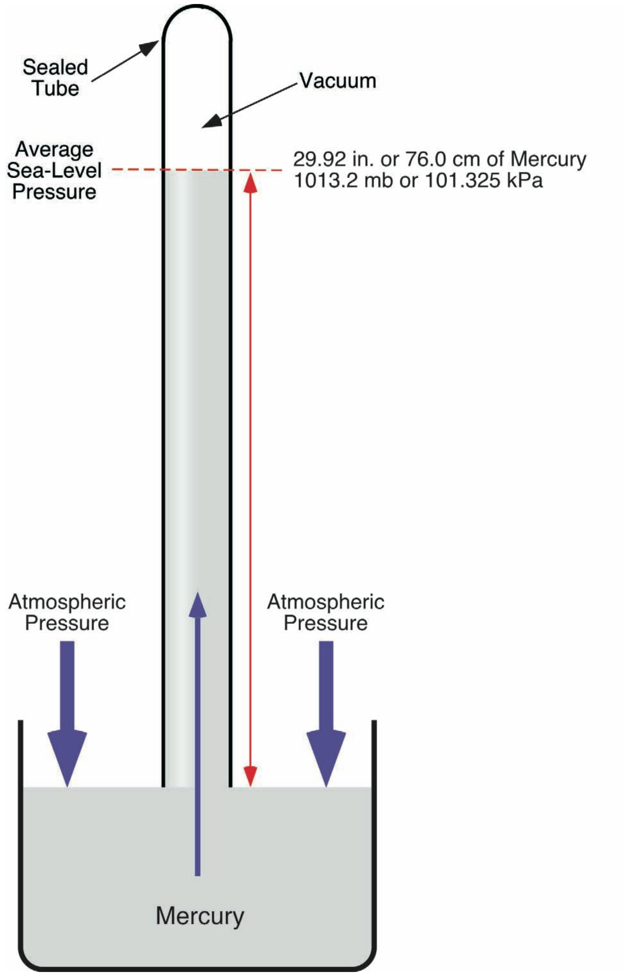


**FIGURE 7.7** Evangelista Torricelli invented the first barometer by immersing a sealed tube in a container of mercury. Pressure from the weight of the overlying atmosphere forced the mercury to move from the container up into the tube to a height of about 30 in. or 76 cm. *(Image Copyright: Michael Pidwirny)*

Torricelli in 1643. In this experiment, Torricelli immersed a tube, sealed at one end, into a container of mercury. Atmospheric pressure then forced the mercury up into the tube to a level that was considerably higher than the mercury in the container (**Figure 7.7**). Torricelli determined from this experiment that the pressure of the atmosphere was about 30 in or 76 cm (1 cm of mercury is equal to 13.3 mb). He also noticed that height of the mercury varied with changes in outside weather conditions.

The most common type of barometer used in homes is the **aneroid barometer** (**Figure 7.8**). The fundamental design of this weather instrument was devised in the 1700s, but was not perfected sufficiently to construct an operational device until the mid-19th century. Inside this instrument is a small, flexible metal capsule called an aneroid cell. In the construction of the device, a partial vacuum is created inside the capsule so that small changes in outside air pressure cause the capsule to expand or contract. The size of the aneroid cell is then calibrated and any change in its volume is transmitted by springs and levers to an indicating arm, which points to the corresponding atmospheric pressure on an indicator display.

A **barograph** is a modified aneroid barometer that is used to continuously record air pressure changes over time (**Figure 7.8**). In this device, the aneroid cell is mechanically linked to a recording pen that produces a continuous record of pressure change with respect to time

**A.**

**B.**

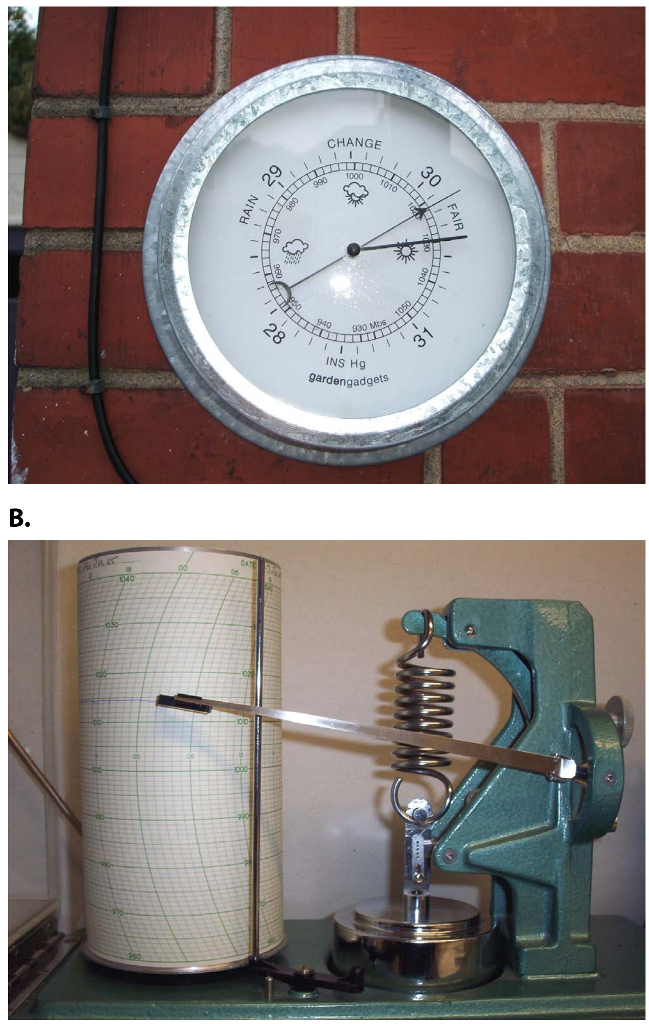


**FIGURE 7.8** (A) Aneroid barometer measuring air pressure in inches and centimeters. (B) Barograph used to obtain continuous records of air pressure change. *(Source: Wikipedia)*

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on a calibrated paper chart. The recording chart, known as a **barogram**, is positioned on a rotating drum that is gradually rotated by a clock mechanism.

## Weather Maps and Air Pressure

Atmospheric pressure readings that are taken routinely at meteorological stations all over the world are used to help construct surface weather maps. This information is shown on these maps using a system of lines called isobars. Isobars are lines connecting points on the map that have the same atmospheric pressure (**Figure 7.9**). **Isobars** are normally drawn on weather maps at standard intervals of 4 mb (996, 1000, 1004, etc.). It is common for these isobars to form an arrangement consisting of a number of enclosed isobars, one inside the other. The centers of such patterns indicate the locations of either high or low-pressure centers on the map. Lows can be identified by the fact that the pressure reading of enclosed isobars decrease as the center of the pattern is approached. Centers of high-pressure display the opposite tendency, the isobars become greater as you move inward.

The isobars drawn on surface weather maps are based on measurements that have been reduced to sea level. Adjustments of air pressure readings to sea level pressure are made regularly at weather stations all across the globe. Essentially, what this adjustment does is remove the influence of elevation from the measurement. As an example, the readings at a location with an elevation of 1000 m (3280 ft) would have 114.4 mb added to them to obtain an equivalent sea level pressure.

## Global Atmospheric Pressure Patterns

**Figure 7.10** describes monthly average sea level pressure for the Earth's surface for January and July. During January, areas of high pressure develop over central Asia (*Siberian High*), off the coast of California (*Hawaiian High*), central North America (*Canadian High*), over Spain and northwest Africa extending into the subtropical North

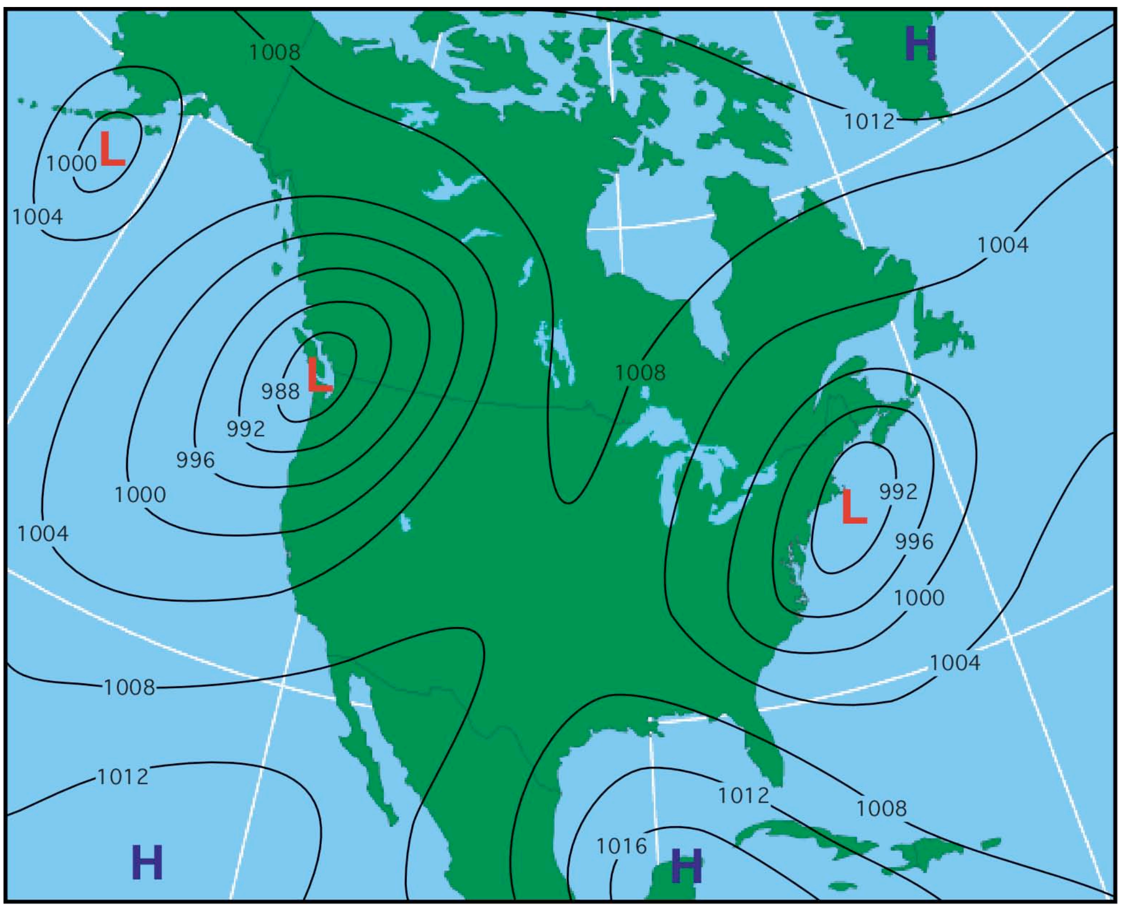


**FIGURE 7.9** Weather map depiction of surface air pressure as measured in millibars (mb). Weather maps use isobars to connect points of equal atmospheric pressure. The interval between isobars shown here is 4 mb. This weather map also indicates centers of low (red L) and high pressure (blue H). *(Image Copyright: Michael Pidwirny)*

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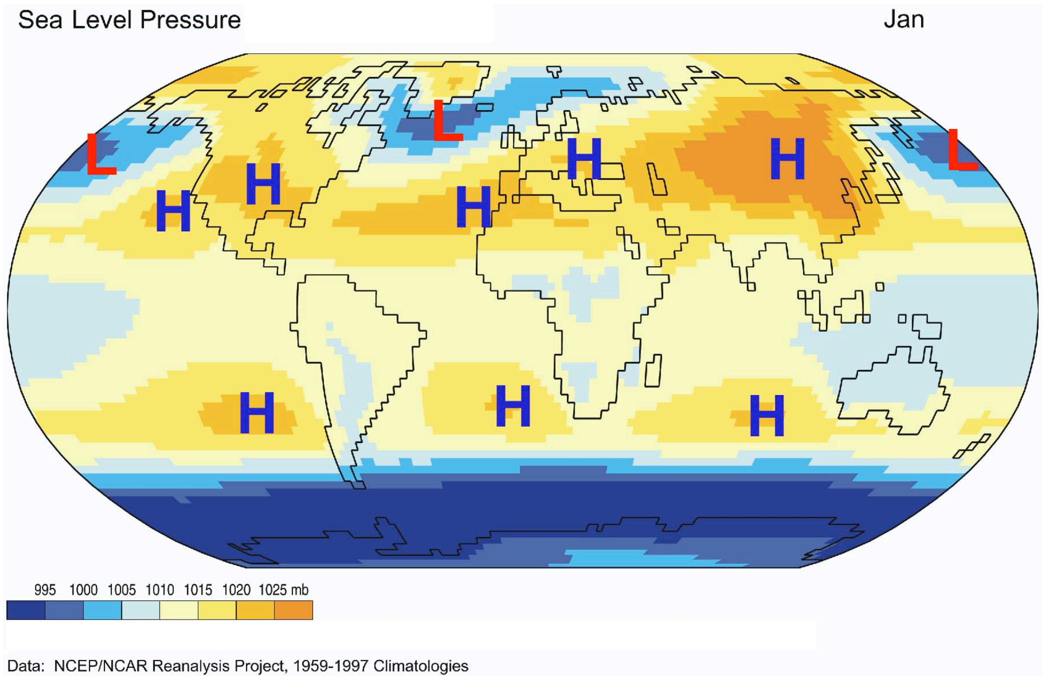
## A. January




## B. July

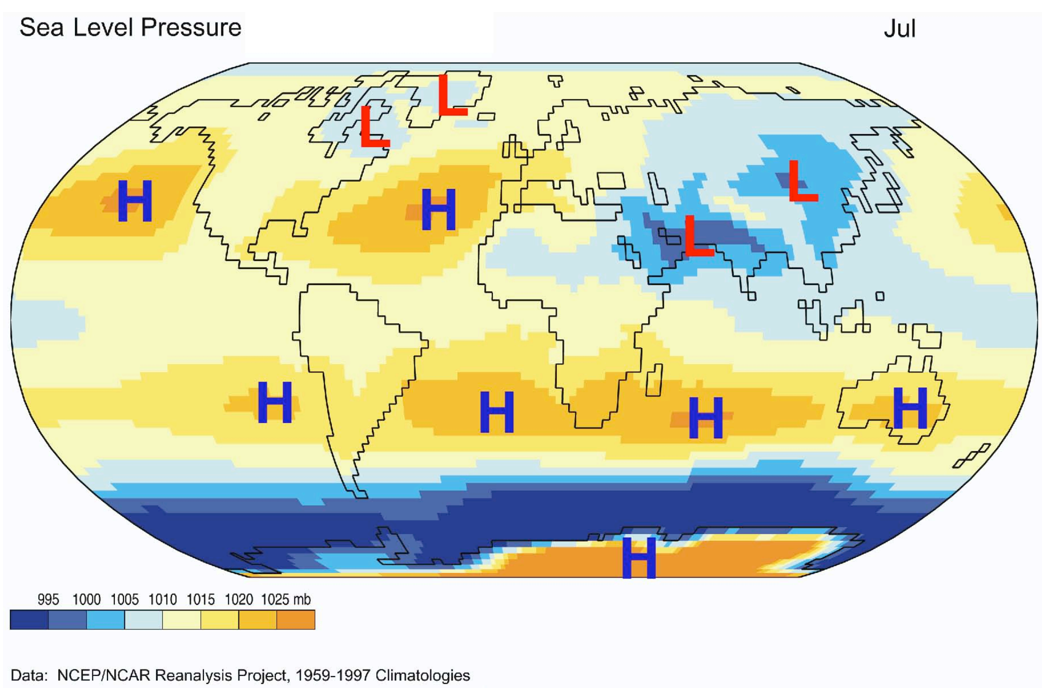



**FIGURE 7.10** January and July average sea level pressure and major centers of low and high pressure on the Earth's surface, 1959-1997. Atmospheric pressure values are adjusted for elevation and are described relative to sea-level. The values are indicated on the map by color shading. Blue shades indicate pressure lower than the global average, while yellow to orange shades are higher than average measurements. *(Original figure courtesy of J.J. Shinker, Department of Geography, University of Oregon).*

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Atlantic (*Azores High*), and over the oceans in the Southern Hemisphere at the subtropics. Areas of low pressure occur just south of the Aleutian Islands (*Aleutian Low*), and at the southern tip of Greenland (*Iceland Low*). We also have a continuous band of low pressure that circles the planet along the coast of Antarctica (50 to 80° South).

During July, a number of dominant winter pressure systems disappear. Gone is the *Siberian High* over central Asia, and the dominant low-pressure systems near the Aleutian Islands and at the southern tip of Greenland. The *Hawaiian* and *Azores High* intensify and expand northward into their relative ocean basins. High pressure systems over the subtropical oceans in Southern Hemisphere also intensity and expand to the north. New areas of dominant high pressure develop over Australia and Antarctica (*South Polar High*). Regions of low pressure form over central Asia and southwest Asia (*Asiatic Low*). These pressure systems are responsible for the summer monsoon rains of Asia. The belt of low pressure found along the coast of Antarctica becomes less intense.

## WIND

**Wind** can be defined simply as air in motion. Near the Earth's surface, the speed of this movement of air can vary from absolute calm to speeds as high as 380 kph (235 mph) (Mt. Washington, New Hampshire, April 12, 1934). Wind can be in any direction, both horizontally (side to side) and vertically (up and down). In most cases, the horizontal component of wind flow greatly exceeds the flow that occurs vertically.

Wind develops as a result of spatial differences in atmospheric pressure. We discovered in the previous section, that changes in air pressure can be driven by a variety of factors, including changes in air temperature (**Figure 7.11**). Consequently, wind speed tends to be at its greatest during the daytime when the greatest spatial extremes in atmospheric temperature and pressure exist.

Wind is often described by two characteristics: wind speed and wind direction. **Wind speed** is the velocity attained by a mass of air traveling horizontally. Wind speed is often measured with an **anemometer** in kilometers per hour (kph), miles per hour (mph), knots, or meters per second (mps) (**Figure 7.12**). **Wind direction** is measured as the direction a wind comes from. For example, a southerly wind comes from the south and blows to the north. A **wind vane** is an instrument used to measure wind direction (**Figure 7.12**). Both of these instruments are positioned in the environment at a standard distance of 10 m above the ground surface.

Winds are named according to the compass direction of their source. Thus, a wind from the north blowing toward the south is called a northerly wind. **Figure 7.13**

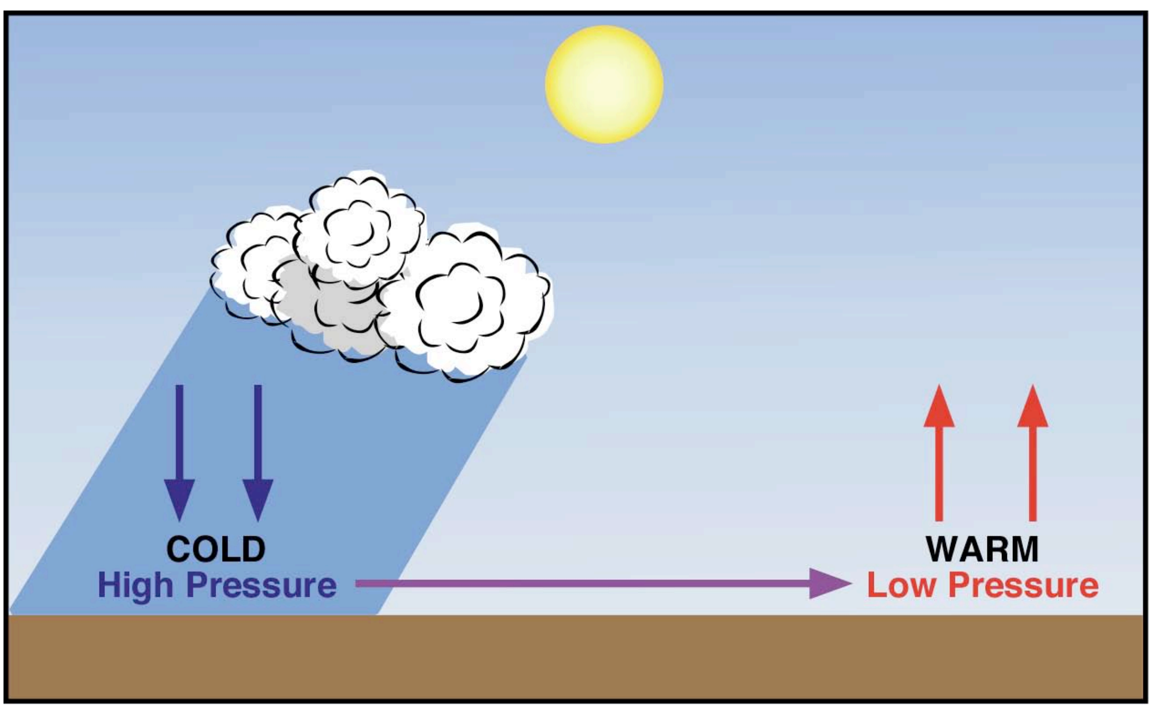


**FIGURE 7.11** Formation of wind as a result of localized temperature differences. *(Image Copyright: Michael Pidwirny)*

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**FIGURE 7.12** Wind speed is commonly measured with an anemometer. An anemometer consists of three open cups attached to a rotating spindle. The speed of rotation is then converted into a measurement of wind speed. Wind direction is measured with a wind vane. On the photograph above, the wind vane instrument has a bullet shaped nose attached to a finned tail by a metal bar. *(Source: NOAA)*

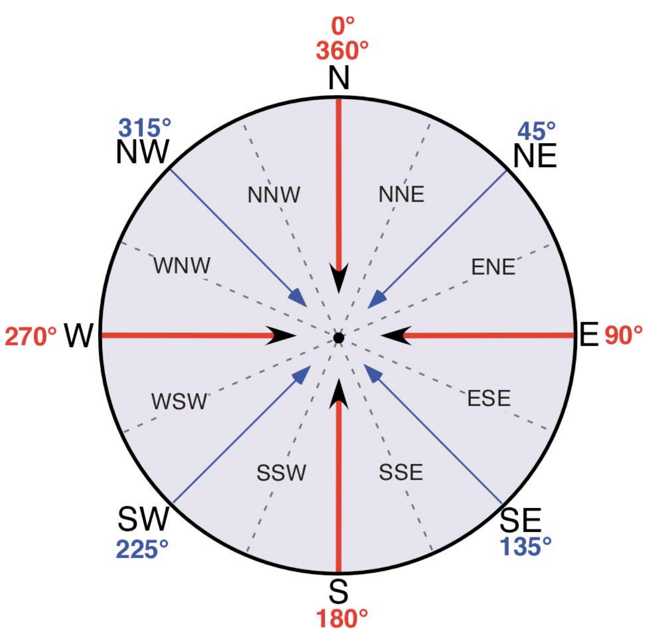


**FIGURE 7.13** Wind compass describing the sixteen principal bearings used to measure wind direction. This system is based on the 360 degrees found in a circle. *(Image Copyright: Michael Pidwirny)*

describes the sixteen principal bearings of wind direction. Most meteorological observations report wind direction using one of these sixteen bearings.

## Newton's Laws of Motion

To understand how wind forms, we must recognize that it is the product of only a few forces. We must also understand that specific fundamental laws of nature control the action of these forces. Sir Isaac Newton, in the year 1670, was the first scientist to describe these laws theoretically. Newton's **first law of motion** suggests that an object that is stationary will remain stationary, and an object that is in motion will stay in motion as long as no opposing force is put on the object. As a result of this law, a puck sent in flight from a blade of a hockey stick will remain in motion until friction slows it down or the goalie makes a save. This law also suggests that once in motion an object's path should be straight, unless it is altered in its direction by another force.

Newton's **second law of motion** suggests that the force put on an object equals its mass multiplied by the acceleration it achieves. The term force in this law refers to the total or net effect of all the forces acting on an object. Mathematically, this law is written as:

$$\text{Force} = \text{Mass} \times \text{Acceleration}$$

or

$$\text{Acceleration} = \text{Force}/\text{Mass}$$

From this natural law of motion, we can see that the acceleration of an object is directly proportional to the net force pushing or pulling that body and inversely proportional to the mass of the body. Thus, the greater the force created by the movement of a hockey player's stick the faster the puck will travel. This law also suggests that if the player used a larger (more massive) puck more force would have to be applied to it to get it to travel as fast as a less massive puck.

We now have a basic understanding of how an object (with mass) will react to a force in terms of motion. We can now apply this knowledge to discovering how a small number of forces are responsible for the creation of wind on our planet.

## Pressure Gradient Force

Horizontally, at the Earth's surface, wind always travels from areas of high pressure to areas of low pressure

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(vertically, winds move from areas of low pressure to areas of high pressure). This situation is comparable to an individual skiing down a hill. The skier will of course move from the top of the hill to the bottom of the hill, with the speed of their descent controlled by the gradient or steepness of the slope. Likewise, wind speed is a function of the steepness or gradient of atmospheric air pressure found between high and low pressure systems. When expressed scientifically, pressure change over a unit distance is known as **pressure gradient force**, and the greater this force, the faster the winds will move along their path.

Pressure gradient force is the primary force influencing the formation of wind from local to global scales. This force is determined by the spatial pattern of atmospheric pressure at any given moment in time. **Figure 7.14** illustrates two different pressure gradient scenarios and their relative effect on wind speed. The two scenarios display the relative relationship between pressure gradient and wind speed. This relationship is linear and positive. As a result, quadrupling the pressure gradient increases wind speed by a factor of four. This is what we would expect according to Newton's second law of motion, assuming the mass of the wind is unchanged.

We can visually see the effect of pressure gradient force on wind speed by examining surface weather maps. On these maps, the relative strength of pressure gradient force can be measured by noting the distance between isobars. If the isobars are closely spaced together, pressure gradient force is great and wind speeds are relatively high (**Figure 7.15**). In areas where the isobars are spaced widely apart, the pressure gradient is low and slower winds will occur.

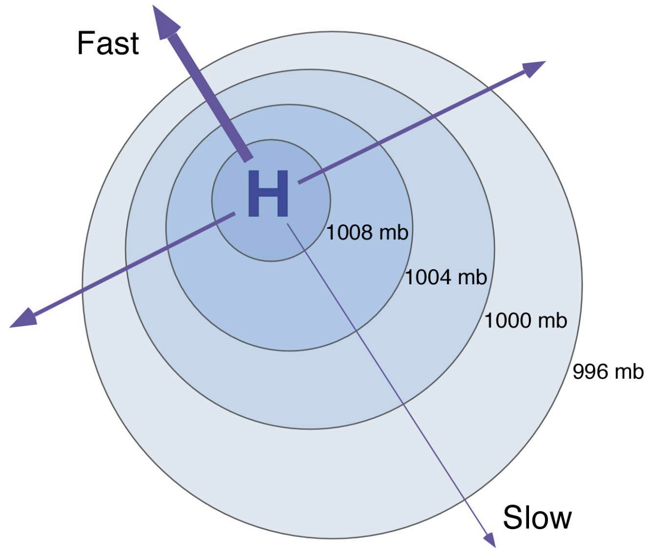


**FIGURE 7.14** Association between wind speed and distance between isobars. In the illustration above, thicker arrows represent relatively faster winds. *(Image Copyright: Michael Pidwirny)*

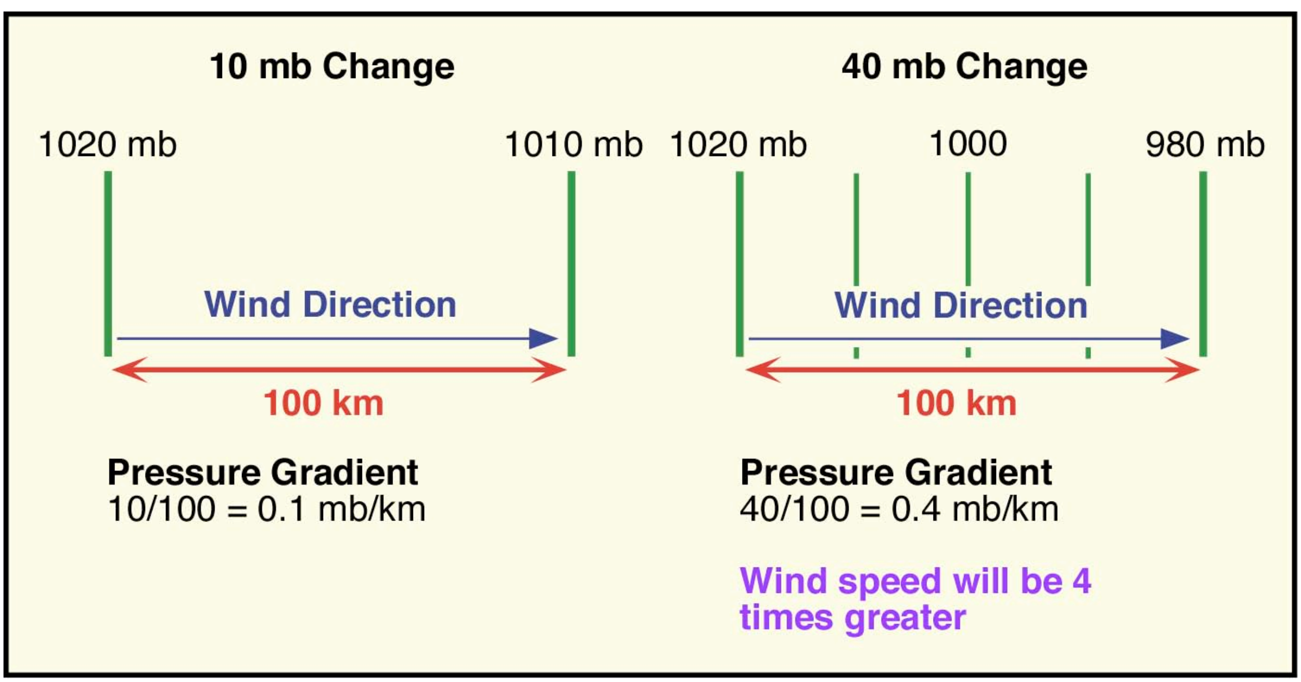


**FIGURE 7.15** Effect of pressure gradient on wind speed. *(Image Copyright: Michael Pidwirny)*

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## Coriolis Effect

In the first half of the 19th century, a French scientist named Gustave-Gaspard Coriolis studied a perplexing problem associated with the movement of winds at the global scale. What he observed was the following: air flowing from high to low pressure systems tends to follow a curved route rather than the shortest straight-line path. He determined that another mechanism was interacting with the movement of the wind causing it to veer of course. He suggested that this additional influence was produced by the rotation of the Earth about its axis. Gustave-Gaspard Coriolis then used an elegant mathematical formula to prove his hypothesis correct. Because of the discovery that Gustave-Gaspard Coriolis made, this *apparent force* that causes winds to deflect off their course is called **Coriolis effect**.

So how does the rotation of the Earth influence the movement of wind? Coriolis effect acts on all free moving objects, including wind, by continually redirecting the path of their motion (Coriolis effect refers to the effect Coriolis force has on a moving object). This redirection occurs because the surface for determining location, the Earth's ground surface, is also moving. Further, this redirection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (**Figure 7.16**). The differences in the direction of this force between the two hemispheres is due to the fact that the Earth's rotation is counter-clockwise north of the equator and clockwise south of the equator (see Figure 4.16). The magnitude of Coriolis effect varies with the velocity and the latitude of the moving object (**Figure 7.17**). Coriolis effect is absent at the equator, and its intensity increases as one approaches either pole.

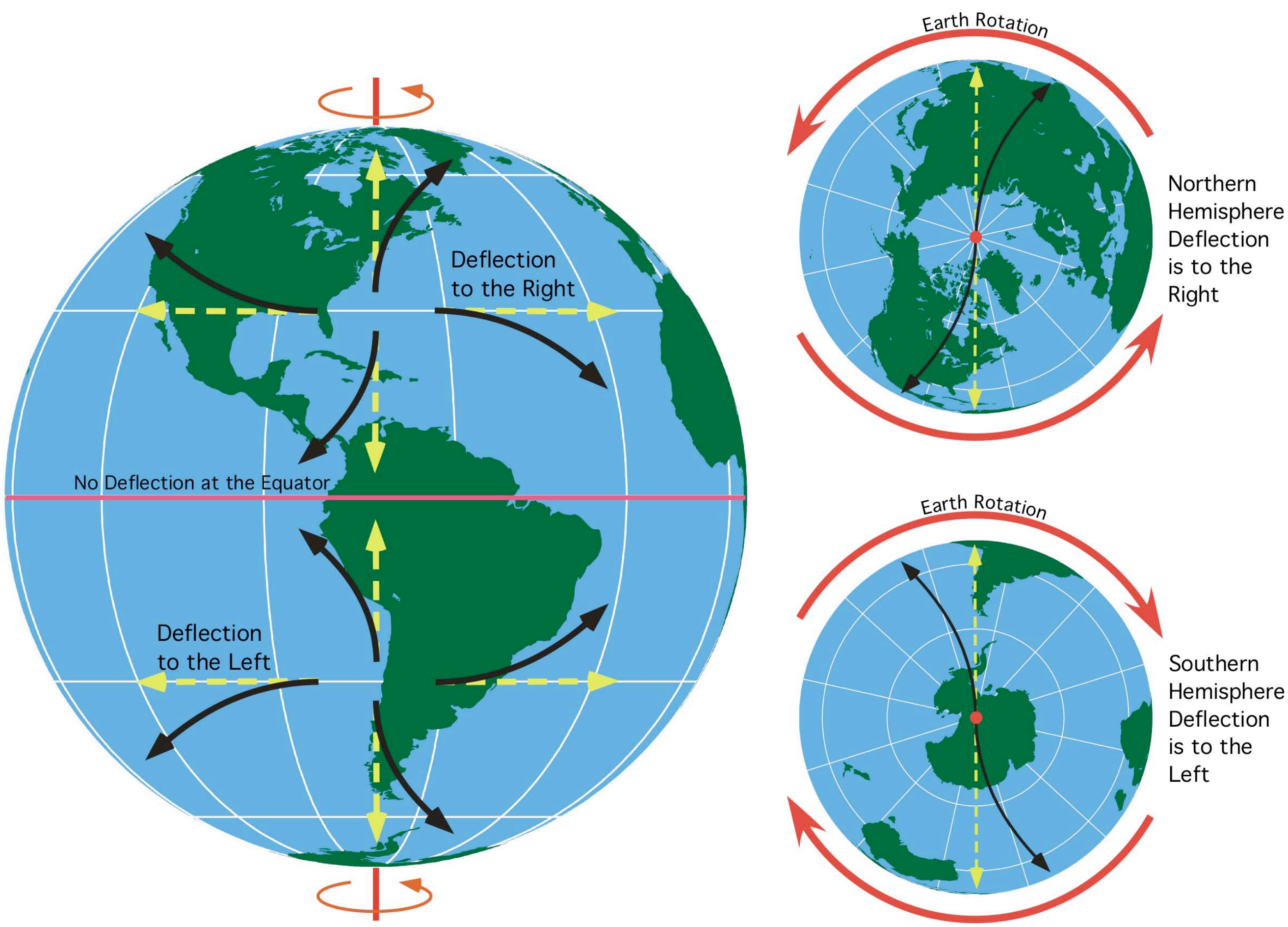


**FIGURE 7.16** The apparent deflection due to the Coriolis effect differs in the North and South Hemisphere. Note that the Coriolis effect does not occur on the Equator. In the Northern Hemisphere objects are deflected to the right, while in the Southern Hemisphere this redirection is to the left. The solid black lines represent the actual route taken by a moving object under the influence of Coriolis effect. The dashed yellow lines represent the intended path. *(Image Copyright: Michael Pidwirny)*

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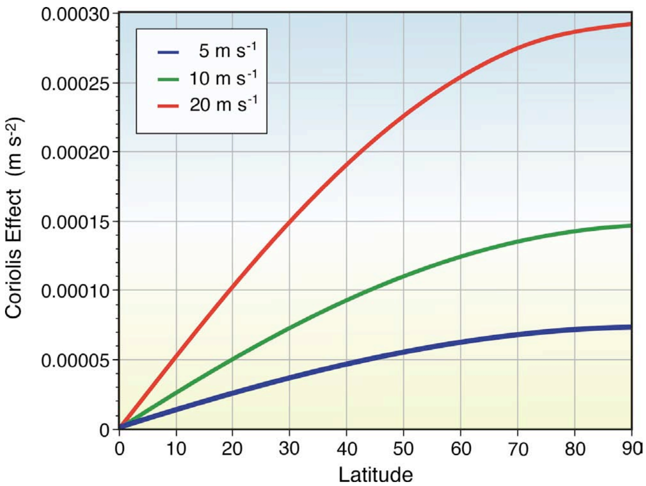


**FIGURE 7.17** Association between wind speed and distance between isobars. In the illustration above, thicker arrows represent relatively faster winds. *(Image Copyright: Michael Pidwirny)*

Additionally, an increase in wind speed also results in a stronger Coriolis effect, and thus in greater deflection of the wind.

The following example may help to illustrate how Coriolis effect works. In **Figure 7.18a**, we graphically model the movement of an airplane from the North Pole to location on the Earth's edge. The destination of the plane is in a straight-line direction along a meridian line. Note that at the same time that the plane is traveling, the Earth's surface is rotating in a counter-clockwise direction. Despite the fact that the airplane is flying in a straight-line, the movement of the Earth's surface makes it appear that the airplane is deviating off course to the right. **Figure 7.18b** shows the same example from the South Pole. Observe that in the Southern Hemisphere the rotation of the Earth is clockwise causing the deflection to be left in direction. **Figure 7.18c** compares the amount of deflection that occurs with latitude. In this figure we can see that the amount of deflection is dependent on latitude. Notice there is no deflection occurring at the equator. For latitudes outside the equator, the magnitude of the deflection increases as one moves closer to the poles.

To complete this discussion on Coriolis effect, we still need to answer one more question. How does Coriolis effect relate to Newtonian laws of motion? According to Newton's first law of motion, air will remain moving in a straight-line unless it is influence by another force. But is Coriolis effect actually pushing or pulling the moving wind in another direction? The answer to this question is no! The Earth's rotation actually exerts no pushing or pulling force on objects moving above the ground surface. In fact, what happens is the ground surface just moves relative to the traveling object making it appear that a second force is pushing (or pulling) the moving object in different direction. We can conclude from this observation that Coriolis effect is not a real force according to Newton's laws of motion. Coriolis effect is an apparent force.

### Frictional Force

Earlier in this section we learned that pressure gradient force is primary force causing wind. Pressure gradient force is also the main force that determines the speed of the wind's movement. Yet, moving air across the Earth's surface can encounter a counteracting force caused by the roughness of the ground surface. This counteracting force is known as **frictional force**.

Friction occurs when an object in motion is hindered or stopped because of physical contact with another object or surface. In the case of wind, friction is caused when objects on the Earth's surface (like houses, trees, mountains, etc) obstruct airflow. The strength of frictional force depends on the types of surfaces that are in contact with wind. Ocean surfaces generally exert less frictional resistance than land. It is also important to note that the interaction of wind with solid obstacles produces swirling masses of air known as **eddies**, another important frictional agent. Eddies can cause opposition to airflow that can extend several hundred meters above the surface that generated them. Because of the effect of eddies and other frictional surfaces, the lower 1000 m (3000 ft) in the atmosphere is often referred to as the **friction layer**. Above this layer, the influence of friction on upper atmosphere airflow is generally negligible.

## SURFACE AND UPPER AIR WINDS

### Surface Winds

Surface winds are the result of three factors: pressure gradient force, Coriolis effect, and frictional force. All surface winds begin with the development of a pressure gradient across space. The force created by the pressure gradient determines initial speed and the preliminary direction of the resulting airflow. On a surface weather map, we can determine the relative strength of pressure gradients at the Earth's surface by examining isobar patterns. The closer the spacing between consecutive isobars the steeper the gradient and the faster the wind will

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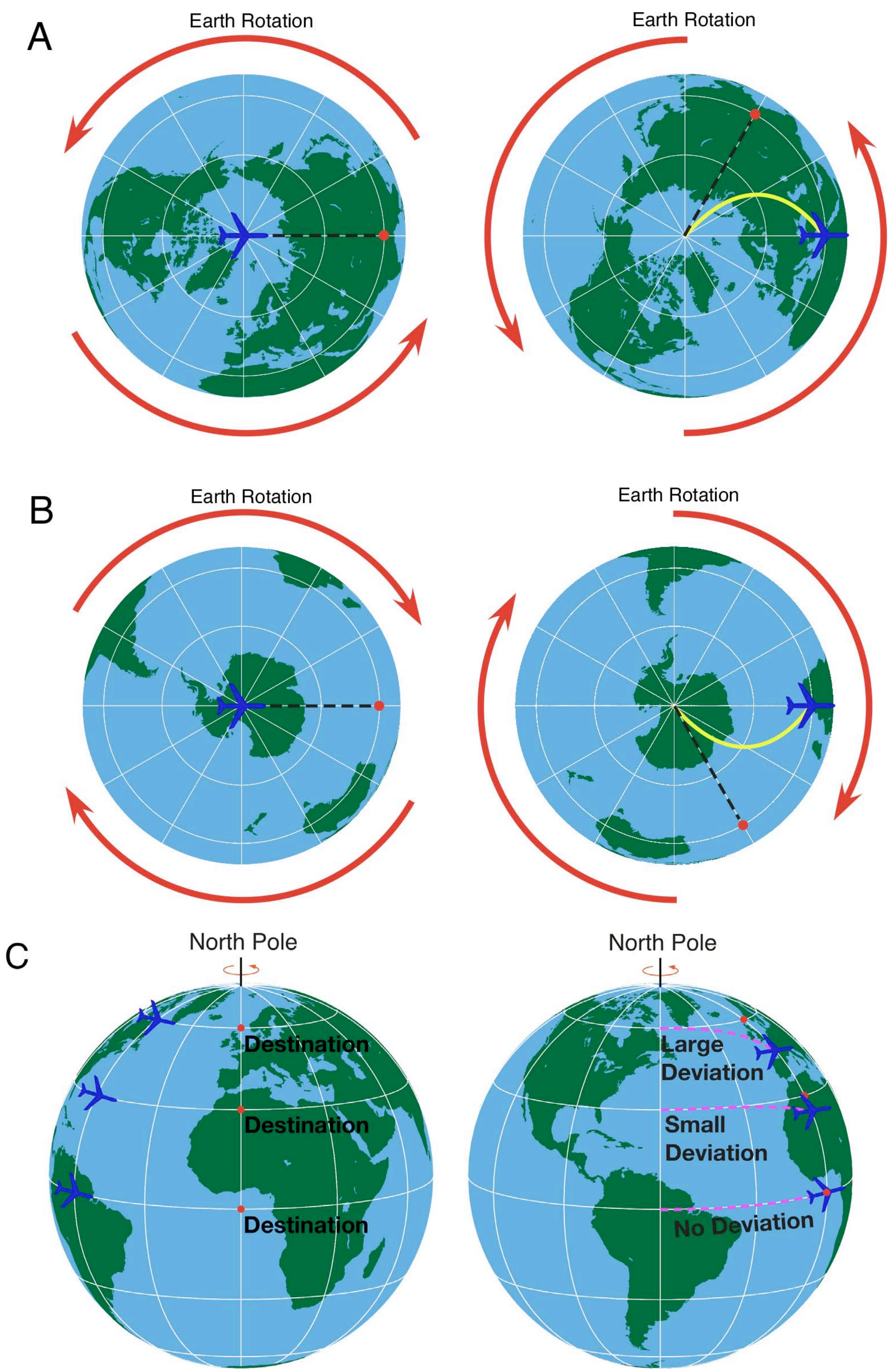


**FIGURE 7.18** Influence of Coriolis effect on moving objects above the Earth's surface. From a perspective above the North Pole, the airplane's path is deflected to the right (A). Above the South Pole, the airplane is deflected from its intended path to the left (B). Illustration C shows how this deviation increases with latitude. Note: at the equator Coriolis effect is zero. *(Image Copyright: Michael Pidwirny)*

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blow. Coriolis effect influences surface winds by deflecting this flow of air off its path. This deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Close to the Earth's surface a third force influences wind: frictional force. Frictional force acts to decelerate moving air. This of course is the opposite of pressure gradient force, which causes acceleration.

**Figure 7.19** graphically models the formation of a surface wind as the combined action of pressure gradient force, Coriolis effect, and frictional force. Each of these factors has a direction relative to the surface wind. As discussed previously, the initial and primary driving force of wind is pressure gradient force. Figure 7.19 indicates that pressure gradient force acts perpendicular to the isobars generating a flow from high to low atmospheric pressure. Coriolis effect acts on wind as soon the air begins to move. Note that this influence is directed to the right (Northern Hemisphere) and at right angles to the direction of the surface wind. Frictional force is shown as a force that acts opposite to direction that surface wind is blowing.

Surface winds tend to cross isobars at angles varying between 10 to 45°. This variation is due to the amount of friction imparted on the moving airflows by the Earth's surface. Less friction means higher wind speeds and greater Coriolis effect. Coriolis effect controls the amount of deflection acting on the wind. In a nearly frictionless environment (like what occurs in the upper atmosphere), Coriolis effect deflects wind until it flows perpendicular to the direction of the pressure gradient force (or parallel to the isobars). Near ground level, the strength of Coriolis effect is reduced because friction decreases wind speed. In turn, this reduction in the Coriolis effect results in less deflection of the surface wind from its path as first defined by pressure gradient force. Over relatively smooth surfaces,

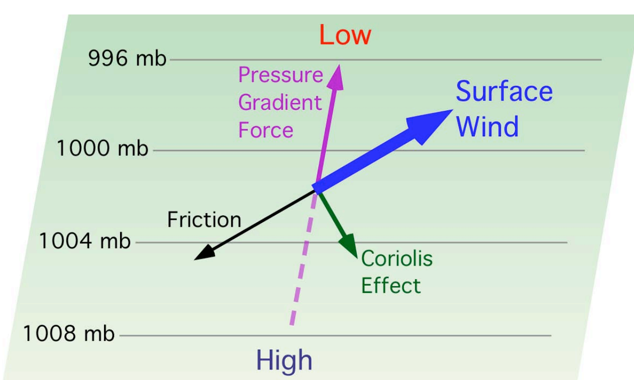


**FIGURE 7.19** Winds near the Earth's surface are influenced by pressure gradient force, Coriolis effect, and surface friction. *(Image Copyright: Michael Pidwirny)*

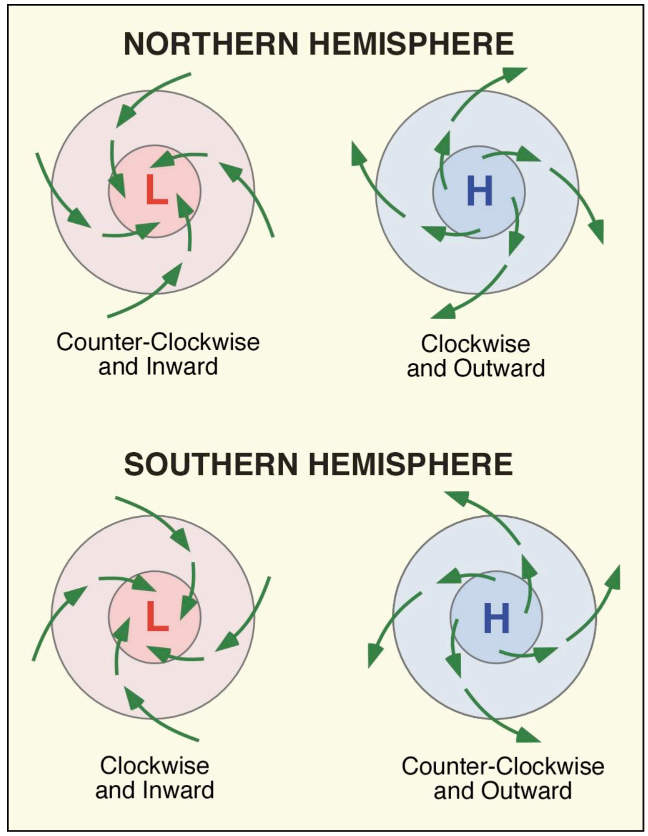


**FIGURE 7.20** Surface wind patterns associated with high and low pressure systems in the Northern and Southern Hemispheres. *(Image Copyright: Michael Pidwirny)*

like oceans, surface winds intersect the isobars at an angle between 10 to 20°. Rough terrestrial surfaces can reduce wind speeds by 30 to 50%. The subsequent reduction in Coriolis deflection causes these winds to intersect isobars at angles between 20 and 45°.

**Figure 7.20** illustrates the patterns of surface airflow that occur around low and high-pressure centers in the Northern and Southern Hemisphere. In the Northern Hemisphere, low-pressure centers (also called **cyclones**) have an airflow that rotates counter-clockwise and inward. Northern Hemisphere high-pressure centers (also called **anticyclones**) have a flow pattern that is outward and clockwise. In the Southern Hemisphere, Coriolis effect acts to the left rather than the right. This causes the winds of the Southern Hemisphere to blow clockwise and inward around surface lows, and counterclockwise and outward around surface highs.

## Upper Air Winds

Earlier in this discussion, we defined upper air winds as any wind that occurs above 1000 m (3000 ft) from the

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Earth's surface. Upper air winds differ from surface winds in one respect: the influence of friction on airflow is minimal. As a result, upper air winds are the product of only two forces: pressure gradient force and Coriolis effect. Figure 7.21 describes the forces acting on both upper air and surface winds. In this figure, we can see that pressure gradient force and Coriolis effect work in opposite directions in an upper air wind. The balance that exists between these two factors also causes an upper air wind to flow parallel to the isobars. Upper air winds flowing in a straight path are commonly called **geostrophic winds**.

## Gradient Winds

Winds above the Earth's surface do not always travel in straight-lines. In many cases, upper air winds flow around the curved isobars of a high (anticyclone) or low (cyclone) pressure centers. Winds that blow around curved isobars above the level of friction are known as **gradient winds**. Figure 7.22 describes the flow patterns of gradient winds around high and low-pressure centers for both the Northern and Southern Hemisphere. Note the difference in the flow direction between the hemispheres.

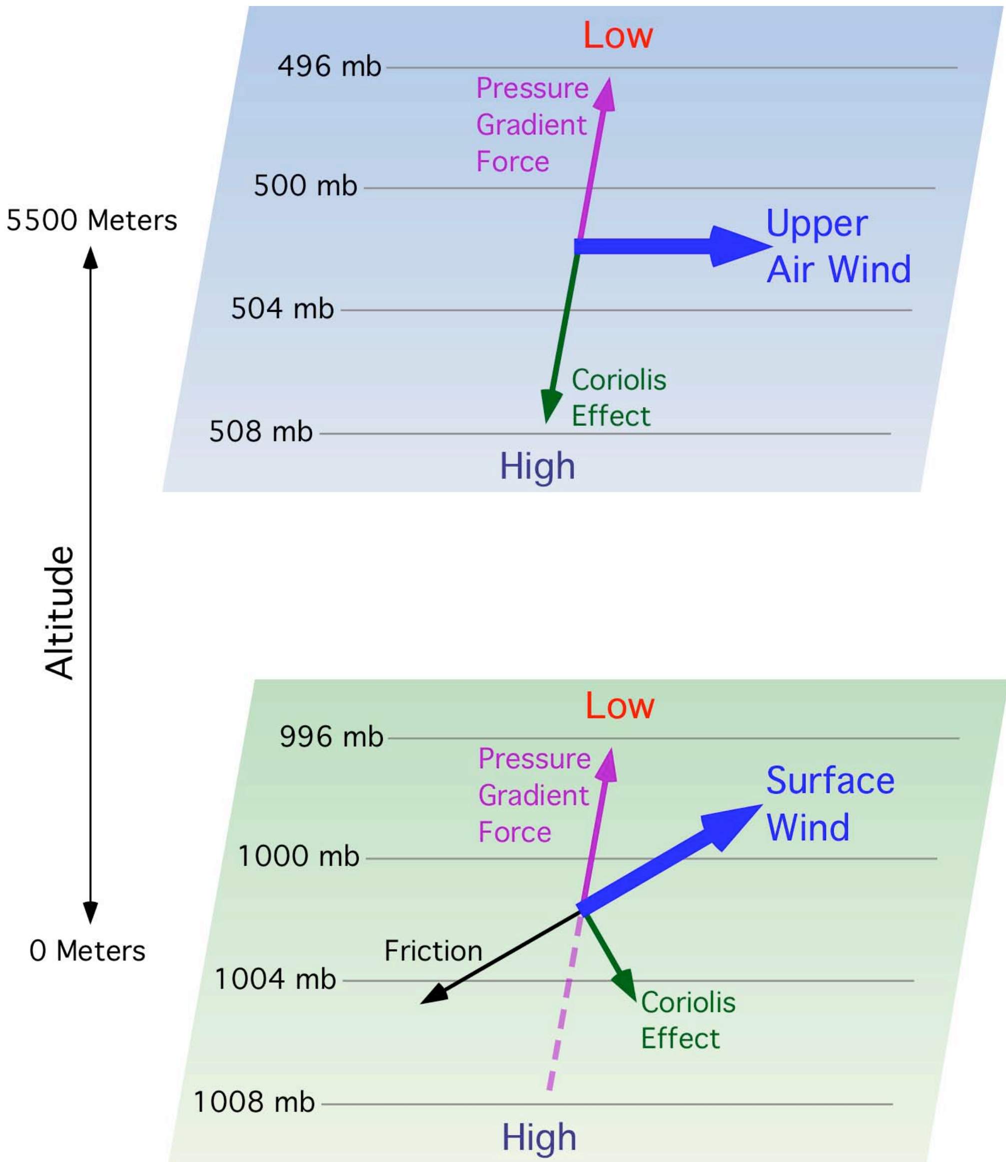


**FIGURE 7.21** Winds near the Earth's surface are influenced by pressure gradient force, Coriolis effect, and surface friction. *(Image Copyright: Michael Pidwirny)*

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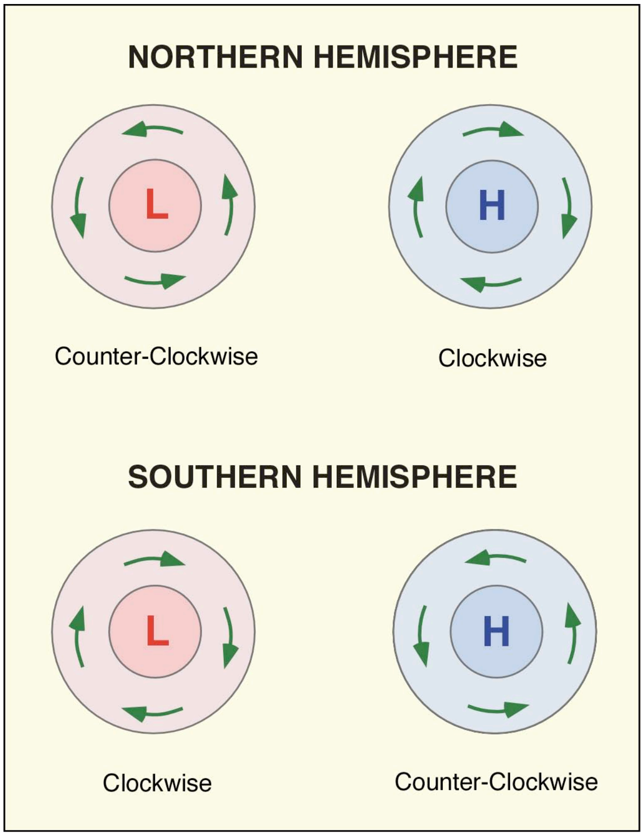


**FIGURE 7.22** Upper atmosphere wind patterns associated with high and low pressure systems in the Northern and Southern Hemispheres. *(Image Copyright: Michael Pidwirny)*

# LOCAL AND REGIONAL WIND SYSTEMS

## THERMAL CIRCULATIONS

As discussed earlier, winds blow because of variations in atmospheric pressure. Pressure gradients may develop on a local to a global scale because of differences in the heating and cooling of the Earth's surface. Heating and cooling cycles that develop daily or annually can create several common local or regional thermal wind systems. The basic circulation system that develops is described in **Figure 7.23**.

In this first diagram (A - **Figure 7.23**), there is no horizontal temperature or pressure gradient and therefore no wind. Atmospheric pressure decreases with altitude as depicted by the drawn isobars (1000 to 980 mb). In the second diagram (B - **Figure 7.23**), the potential for solar heating is added which creates contrasting surface areas of temperature and atmospheric pressure. The area to the right receives more solar radiation and the air begins to warm from heat energy transferred from the ground through conduction and convection. The vertical distance between the isobars becomes greater as the air rises. To the far left, less radiation is received because of the presence of cloud, and this area becomes relatively cooler than the area to the right. In the upper atmosphere, a pressure gradient begins to form because of the rising air and upward spreading of the isobars. The air then begins to flow in the upper atmosphere from high pressure to low pressure.

Diagram C (**Figure 7.23**) shows a fully developed thermal circulation system. Beneath the upper atmosphere high, is a thermal low-pressure center created from the heating of the ground surface. Below the upper atmosphere low, is a thermal high created by the relatively cooler air temperatures and the descend air from above. Surface air temperatures are cooler here because of the obstruction of shortwave radiation absorption at the Earth's surface by the cloud. At the surface, the wind blows from the high to the low pressure. Once at the low, the wind rises up to the upper air high-pressure system because of thermal buoyancy and outflow in the upper atmosphere. From the upper high, the air travels to the upper air low, and then back down to the surface high to complete the circulation cell. Note that the thermal circulation cell is a closed system that redistributes air from areas that have a surplus to areas that have a deficit (in terms of air pressure). This circulation cell is driven by the greater heating of the surface air in the right side of the diagram.

## SEA AND LAND BREEZES

Sea and land breezes are types of thermal circulation systems that develop at the interface of land and ocean or large bodies of water. At this interface, the dissimilar heating and cooling characteristics of land and water initiate the development of an atmospheric pressure gradient that causes the air in these areas to flow. During the daytime land heats up much faster than water as it receives solar radiation from the Sun (**Figure 7.24**). The warmer air over the land then begins to expand and rise forming a thermal low. At the same time, the air over the ocean becomes a cool high because of water's slower rate of heating. Air begins to flow as soon as there is a significant difference in air temperature and pressure across the land to sea gradient. The development of this pressure gradient causes the heavier cooler air over the ocean to move toward the land and to replace the air rising in the thermal low. This localized airflow system is called a **sea breeze**. Sea breeze usually begins in midmorning and reaches its maximum strength in the later afternoon when

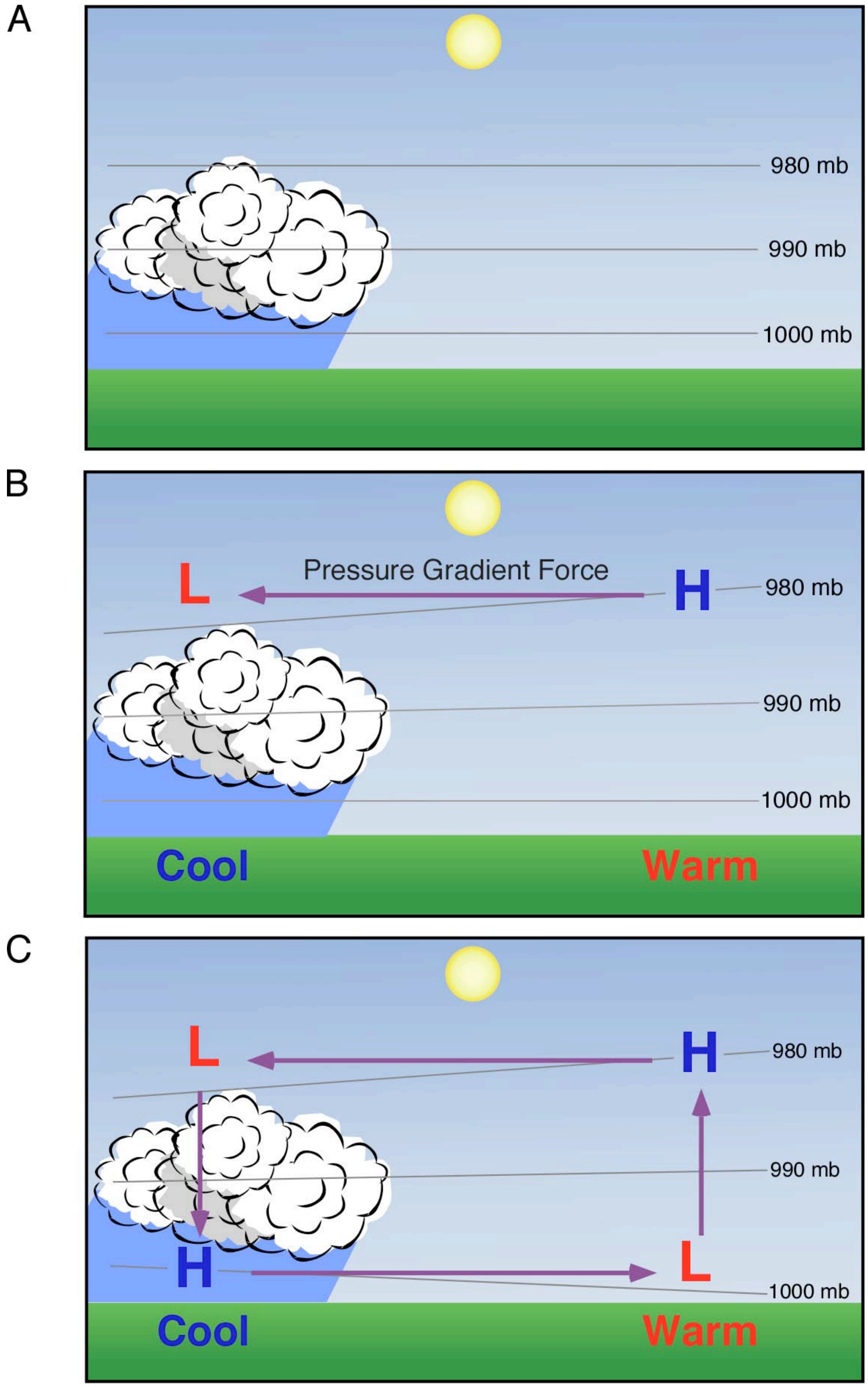


**FIGURE 7.23** The formation of a thermal circulation system. This process begins with the unequal heating of the Earth's surface and lower atmosphere (A). The unequal heating causes the isobars over the area being heated to spread apart because of convection and air expansion. This process also initiates the horizontal flow of air in the upper atmosphere (B). Air flow patterns then begin in the vertical and near the ground surface completing the circulation system (C). *(Image Copyright: Michael Pidwirny)*

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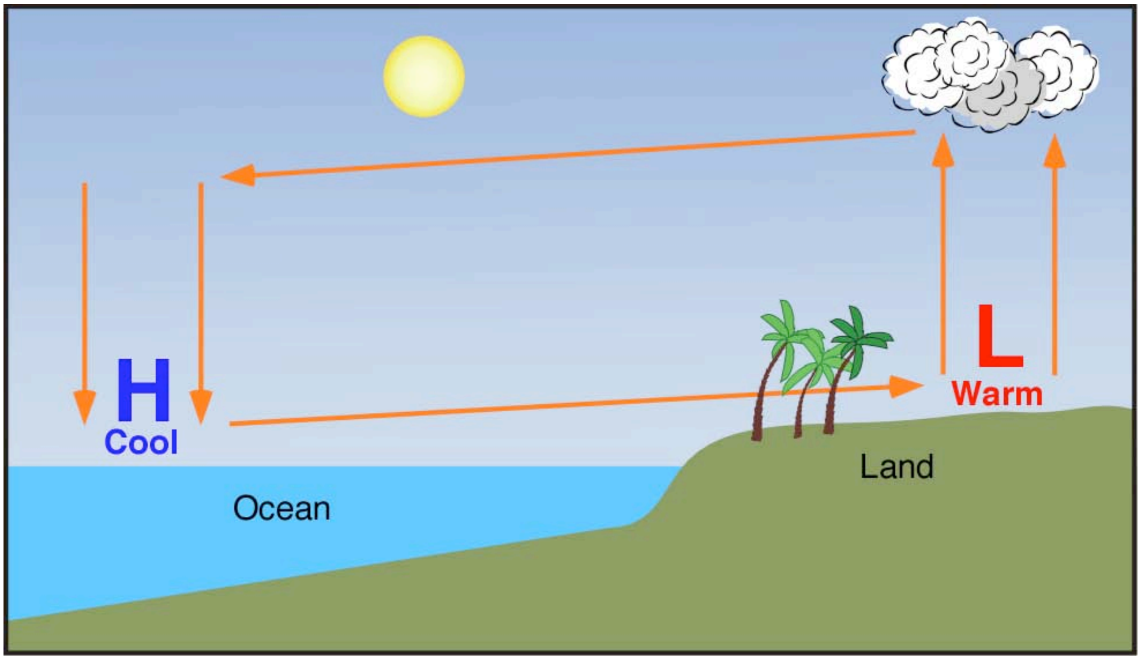


**FIGURE 7.24** Daytime development of sea breeze. *(Image Copyright: Michael Pidwirny)*

the greatest temperature and pressure contrasts exist. Sea breeze dies down at sunset when air temperature and pressure once again become similar across the two surfaces.

At sunset, the land surface stops receiving radiation from the Sun (**Figure 7.25**). As night continues, the land surface begins losing heat energy at a much faster rate than the water surface. After a few hours, air temperature and pressure contrasts begin to develop between the land and ocean surfaces. The land surface being cooler than the water becomes a thermal high-pressure area. The ocean becomes a warm thermal low. Wind flow now moves from the land to the open ocean. This type of localized airflow is called a **land breeze**.

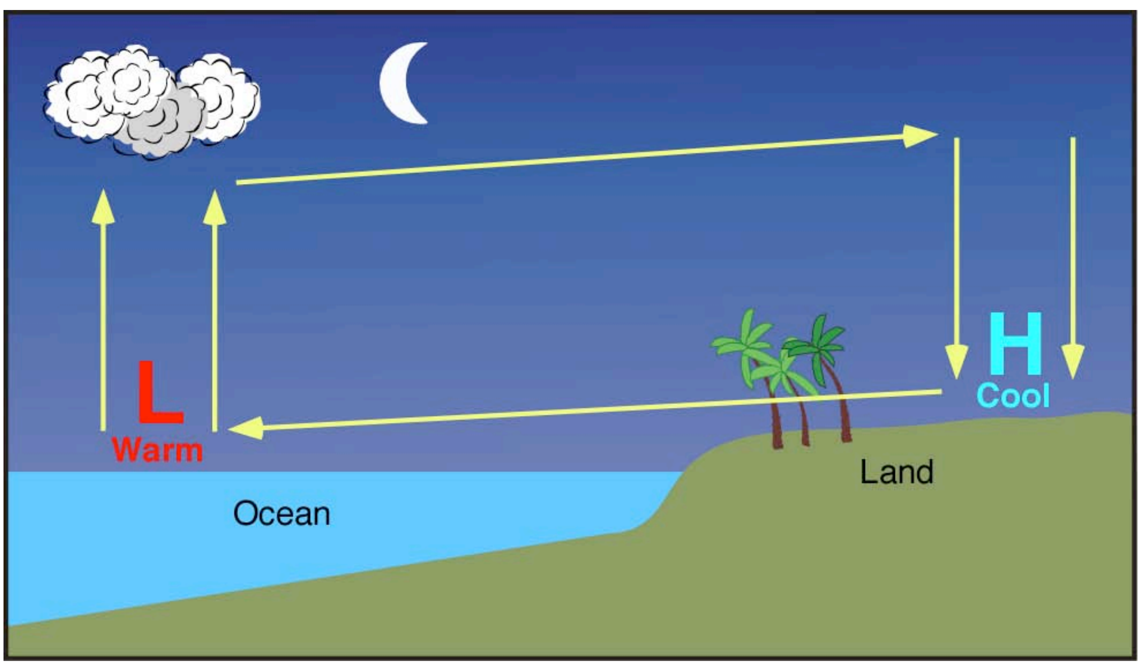


**FIGURE 7.25** Night time development of land breeze. *(Image Copyright: Michael Pidwirny)*

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## Mountain and Valley Breezes

Mountain and valley breezes are common in regions with great topographic relief. A **valley breeze** develops during the day as the Sun heats the land surface and air at the valley bottom and sides (**Figure 7.26**). As the air heats it becomes less dense and buoyant and begins to flow gently up the valley sides. Vertical ascent of the air rising along the sides of the mountain is usually limited by the presence of a temperature inversion layer. When the ascending air currents encounter the inversion they are forced to move horizontally and then back down to the valley floor. This creates a self-contained circulation system. If conditions are right, the rising air can condense and form into cumuliform clouds.

During the night, the air along the mountain slopes begins to cool quickly because of longwave radiation loss (**Figure 7.27**). As cooling proceeds through the night, the air at high elevations becomes denser and begins to flow downslope because of gravity creating a **mountain breeze**. Convergence of the draining air occurs at the valley floor and forces the air to move vertically upward. The upward movement is usually limited by the presence of a temperature inversion that forces the air to begin moving

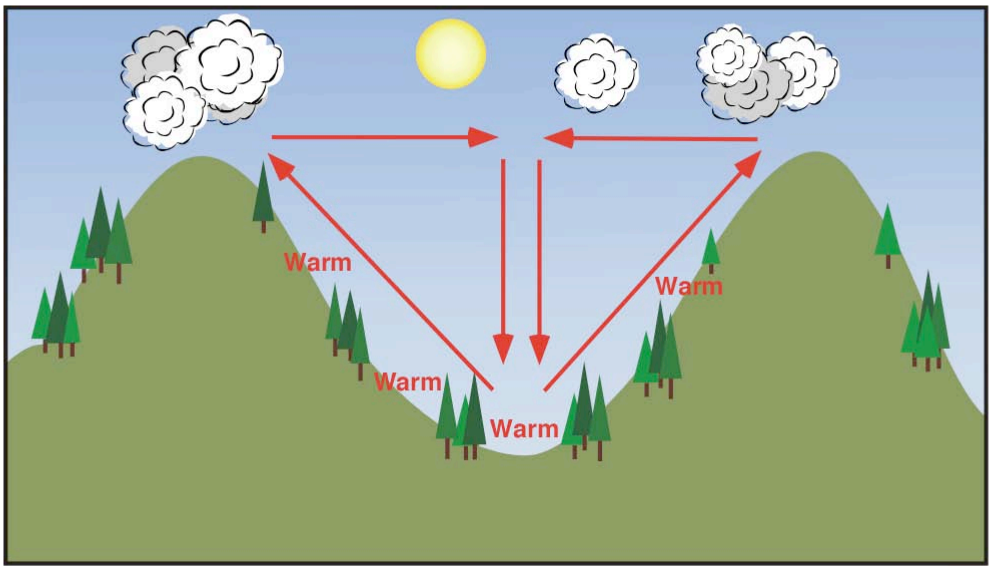


**FIGURE 7.26** Daytime development of valley breeze. *(Image Copyright: Michael Pidwirny)*

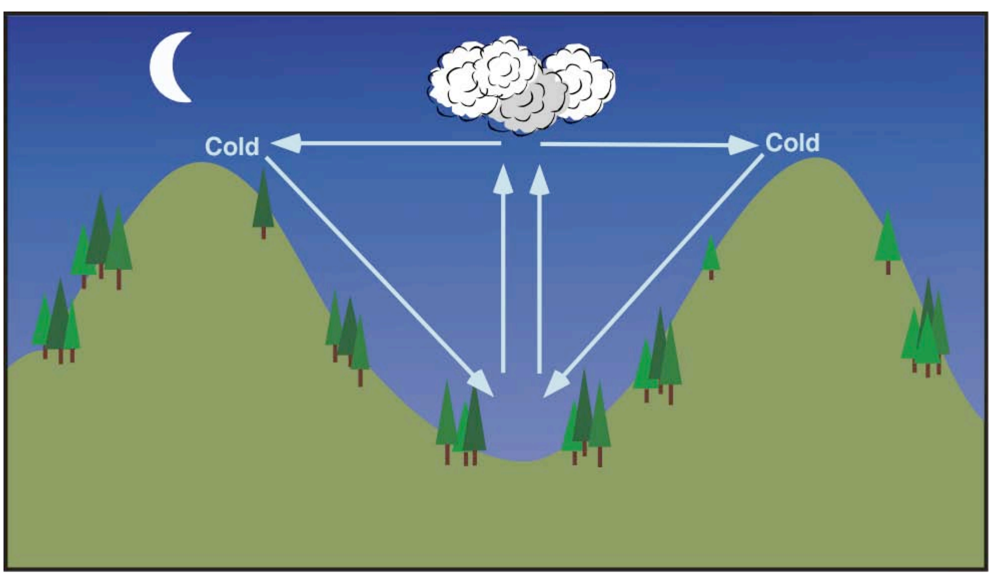


**FIGURE 7.27** Night time development of mountain breeze. *(Image Copyright: Michael Pidwirny)*

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horizontally. This horizontal movement completes the circulation cell system.

### Katabatic Winds

**Katabatic winds** are a special type of mountain breeze that have relatively faster flow velocities. Katabatic winds develop under a specific set of environmental conditions. These environmental conditions include: the presence of an extensive elevated plateau and sufficient cooling of a large air mass by longwave radiation loss. Katabatic winds begin flowing when the developing pool of cold air becomes large enough to be influenced by gravity and topography. The flow stops when all of the pooled cold air drains from the plateau to locations downslope. Some katabatic winds can quickly accelerate in speed if the mass of air is forced to flow through narrowing terrain. This type of acceleration process is known as a **venturi effect**.

Katabatic winds are very common along the glacial ice sheet margins of Greenland and Antarctica. Cape Denison, Antarctica can experience katabatic winds with velocities as high as 200 kph (120 mph). This location also holds the title of having the highest recorded average wind speed on our planet.

### Monsoon Winds

**Monsoons** are continental scale wind systems that predictably change direction with the passing of the seasons. Like land/sea breezes, these wind systems are created by the temperature contrasts that exist between the surfaces of land and ocean. Monsoons are different from land/sea breezes both spatially and temporally. Monsoons occur over areas of thousands of square kilometers, with an annual cycle period.

During the summer, monsoon winds blow from the cooler ocean surfaces onto the warmer continents (**Figure 7.28**). In the summer, the continents become much warmer than the oceans because of a number of factors. These factors include:

- Specific heat differences between land and water.
- Greater evaporation over water surfaces.
- Subsurface mixing in ocean basins that redistributes heat energy through a deeper layer.

Precipitation is normally associated with the summer monsoons (**Figure 7.29**). Onshore winds blowing inland from the warm ocean are very high in humidity, and slight cooling of these air masses causes condensation and rain. In some cases, this precipitation is greatly intensified by **orographic uplift**. Some highland areas in Asia receive more than 10 m (33 ft) of rain during the summer months.

In the winter the wind patterns reverse because the ocean surfaces are now warmer. With less solar energy available, the continents begin cooling rapidly as longwave radiation is emitted to space. The ocean surface retains its heat energy longer because of water's high specific heat and subsurface mixing. The winter monsoons bring clear dry weather and winds that flow from land to sea.

**Figure 7.28** illustrates the general wind patterns associated with the winter and summer monsoons in Asia. The Asiatic monsoon is the result of a complex climatic interaction between the distribution of land and water, topography, and tropical and mid-latitudinal circulation. In the summer, a low-pressure center forms over northern India and northern Southeast Asia because of higher solar insolation. Warm moist air is drawn into the thermal lows from air masses over the Indian Ocean. Summer heating also causes the development of a strong latitudinal pressure gradient and the development of an easterly jet stream at an altitude of about 15 km (9.3 mi) and a latitude of 25° North. The jet stream enhances rainfall in Southeast Asia, in the Arabian Sea, and in South Africa. When autumn returns to Asia the thermal extremes between land and ocean decrease and the Westerlies of the mid-latitudes move in. The easterly jet stream is replaced with strong westerly winds in the upper atmosphere. Subsidence from an upper atmosphere cold low above the Himalayas produces outflow that creates a surface high-pressure system that dominates the weather in India and Southeast Asia. Monsoon wind systems also exist in Australia, Africa, South America, and North America.

## MODELS OF GLOBAL CIRCULATION

We can gain an understanding of how global circulation works by developing two visual models of the processes that produce this system. The first model will be founded on the following simplifying assumptions:

- The Earth is not rotating on its axis in space.
- The Earth's surface is composed of the same material throughout.

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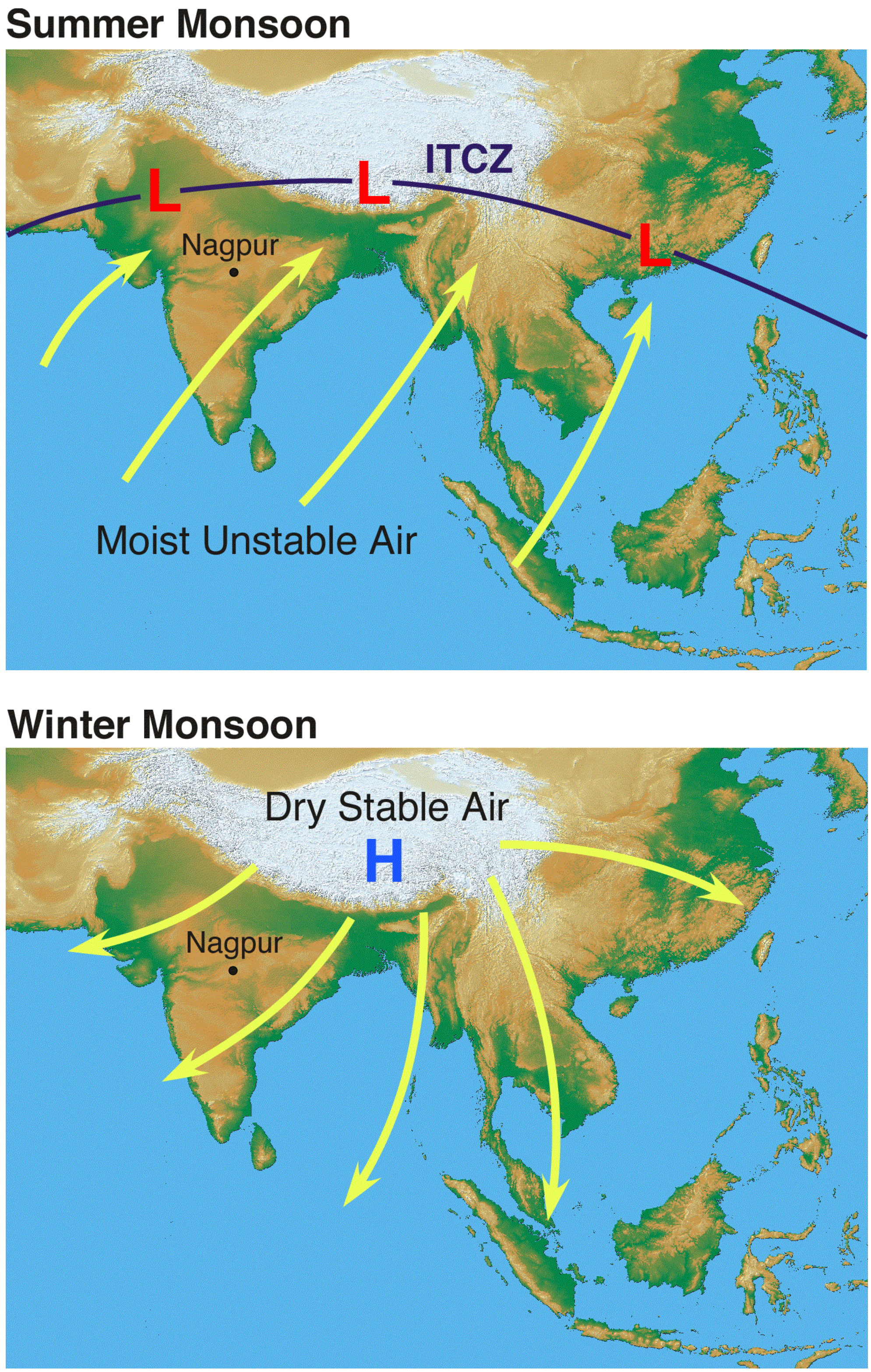


**FIGURE 7.28** Summer and winter monsoon wind patterns for Southeast Asia. Winds blow from the Indian Ocean to the Asian landmass during the summer monsoon. The air over the Indian Ocean is warm and very moist. When this air reaches the continent, changes in elevation cause the development of precipitation because of orographic uplift. During the winter monsoon, the winds blow from the Asian continent to the oceans. This air is dry and stable. . *(Image Copyright: Michael Pidwirny)*

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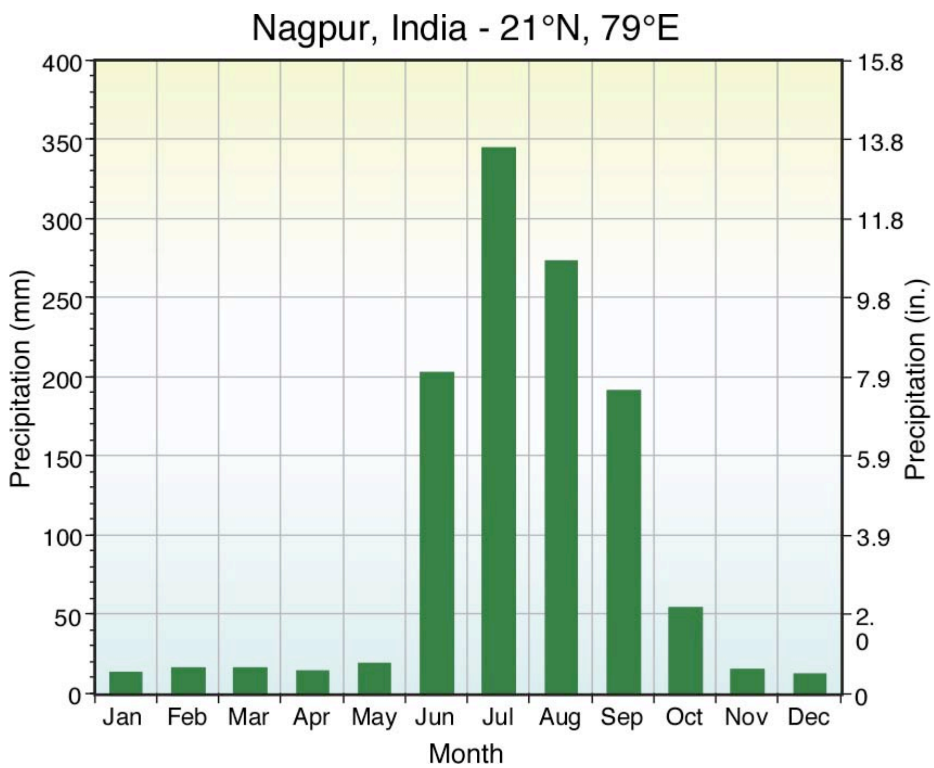


**FIGURE 7.29** Monthly precipitation at Nagpur, India shows a summer maximum enhanced by monsoon circulation. The summer monsoon draws warm moist air from the Indian Ocean to Nagpur during the months of June, July, August, and September. *(Image Copyright: Michael Pidwirny)*

- The global reception of solar insolation and loss of longwave radiation cause a temperature gradient of hotter air at the equator and colder air at the poles.

Based on these assumptions, air circulation on the Earth should approximate the patterns shown on **Figure 7.30**. In this illustration, each hemisphere contains one three-dimensional circulation cell.

As described in the diagram above, surface airflow is from the poles to the equator. When the air reaches the equator, the processes of convection and convergence act together to cause vertical uplift. The air over the equator continues to rise until it reaches the top of the troposphere. At the top of the troposphere, the air begins to flow once again horizontally toward the poles. At the poles, the air in the upper atmosphere then descends to the Earth's surface to complete the cycle of flow.

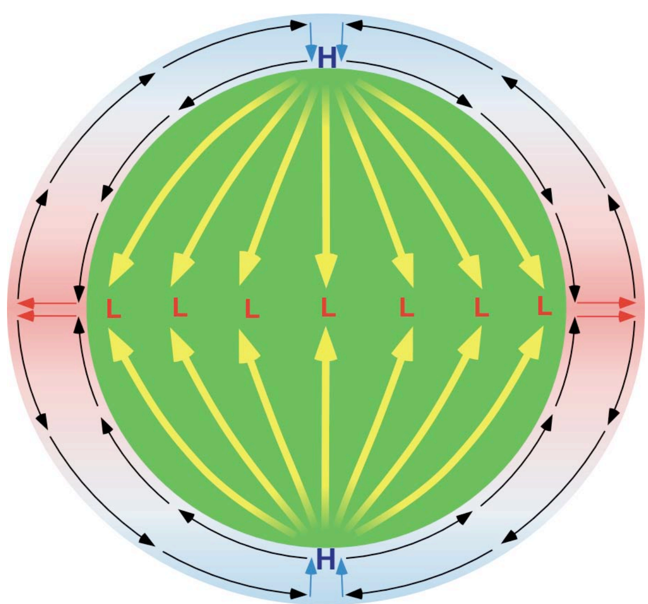


**FIGURE 7.30** Simplified one-cell global air circulation patterns. *(Image Copyright: Michael Pidwirny)*

## Three Cell Model of Global Circulation

If we eliminate the first assumption, the pattern of flow described in the simple model above would be altered, and the circulation of the atmosphere would more closely resemble the actual global patterns found on the Earth. Planetary rotation would cause the development of three circulation cells in each hemisphere rather than one single circulation cell (**Figure 7.31**). These three circulation cells are known as the: **Hadley Cell**; **Ferrel Cell**; and **Polar Cell**.

In the new model, the equator still remains the warmest location on the Earth. This area of greater heat acts as a zone of thermal lows known as the **intertropical convergence zone** (ITCZ). The ITCZ draws in surface air from the subtropics. When this subtropical air reaches the equator, it rises into the upper atmosphere because of **convergence** and **convection**. It attains a maximum vertical altitude of about 16 km (9.9 mi). This height represents the top of the troposphere. At the top of the troposphere, the rising air changes direction and begins traveling horizontally towards the North and South Poles. Coriolis effect causes the deflection of this moving air in the upper atmosphere, and by about 30° of latitude the air begins to flow zonally from west to east. This zonal flow is known as the **subtropical jet stream**. The zonal flow also causes the accumulation of air in the upper atmosphere, as it is no longer flowing **meridionally**. To compensate for this accumulation, some of the air in the upper atmosphere sinks back to the surface creating the **subtropical high-pressure zone**. From this zone, the surface air travels in two directions. A portion of the air moves back toward the

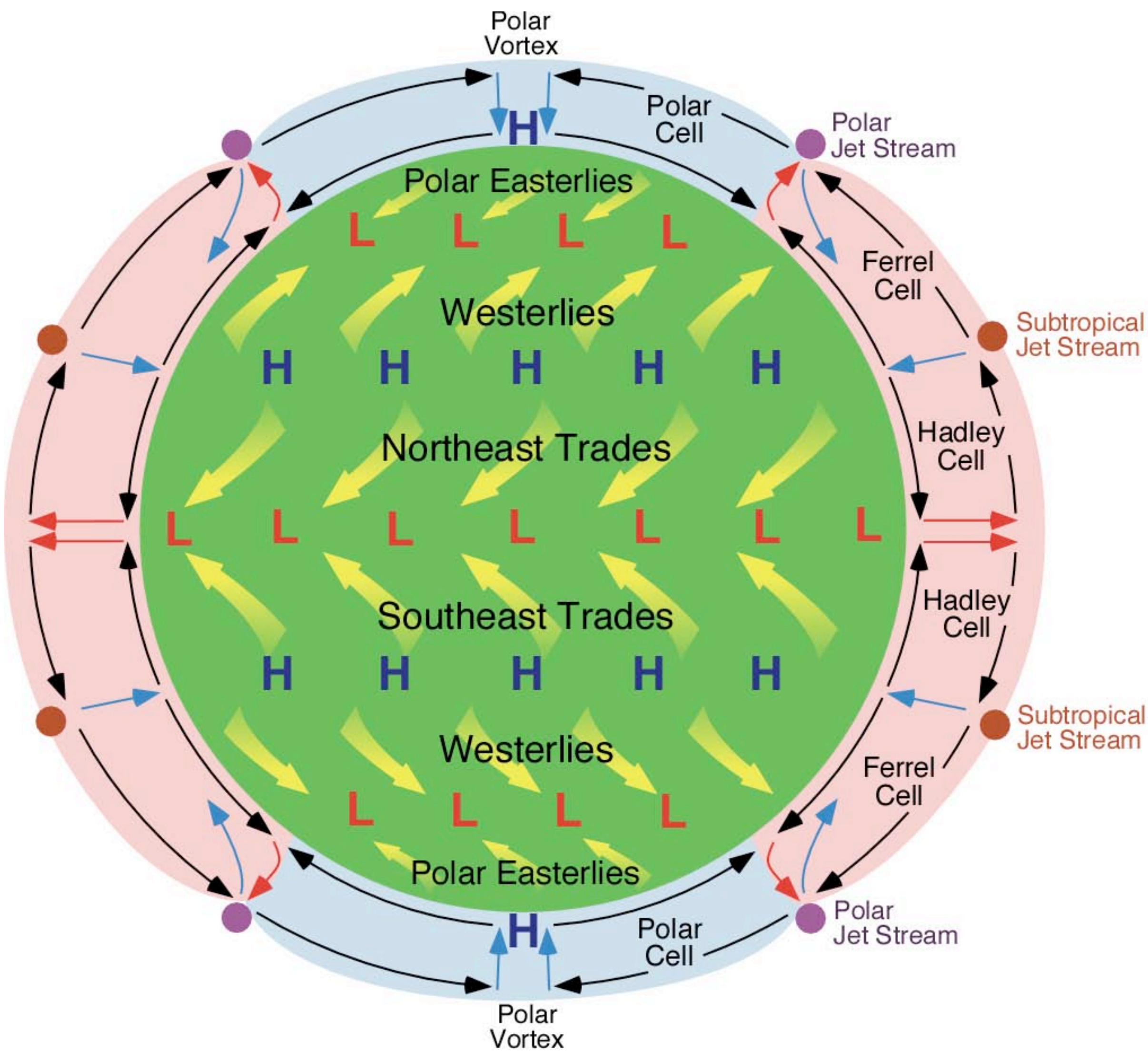


**FIGURE 7.31** Idealized three-cell global circulation model *(Image Copyright: Michael Pidwirny)*

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equator completing the circulation system known as the Hadley cell. This moving air is also deflected by the Coriolis effect to create the **Northeast Trades** (right deflection) and **Southeast Trades** (left deflection). The surface air moving towards the poles from the subtropical high zone is also deflected by Coriolis effect producing the **Westerlies**. Between the latitudes of 30 to 60° North and South, upper air winds blow generally towards the poles. Once again, Coriolis effect deflects this wind causing it to flow in a west to east at roughly 60° North and South. This zone of high-speed upper atmosphere wind is called **polar jet stream**. On the Earth's surface at 60° North and South latitude, the Westerlies collide with cold air traveling from the poles. This collision results in frontal uplift and the creation of the **polar front** and **subpolar lows** (**mid-latitude cyclones**). A small portion of this lifted air is sent back into the Ferrel cell after it reaches the top of the troposphere. Most of this lifted air is directed to the **polar vortex** where it moves downward to create the **polar high**.

## ACTUAL GLOBAL SURFACE CIRCULATION

**Figure 7.32** describes the actual surface circulation for the Earth as determined from 39 years of record. Average circulation patterns can be viewed on this animation for every month of the year. The circulation patterns seen differ somewhat from the three cell model in **Figure 7.31**. These differences are caused primarily by two factors. First, the Earth's surface is not composed of uniform materials. The two surface materials that dominate are water and land. These two materials behave differently in terms of heating and cooling causing latitudinal pressure zones to be less uniform. The second factor influencing actual circulation patterns is elevation. Elevation tends to cause pressure centers to become intensified when altitude is increased. This is especially true of high-pressure systems.

**Figure 7.32** displays average surface wind patterns for January and July. The intertropical convergence zone (ITCZ) is identified on the figures by a red line. The formation of this band of low pressure is the result of solar heating and the convergence of the trade winds. In January, the intertropical convergence zone is found south of the equator (**Figure 7.32A**). During this time period, the Southern Hemisphere is tilted towards the Sun and receives higher inputs of shortwave radiation. Note that the line representing the intertropical convergence zone is not straight and parallel to the lines of latitude. Bends in the line occur because of the different heating characteristics of land and water. Over the continents of Africa, South America, and Australia, these bends are toward the South Pole. This phenomenon occurs because land heats up faster than the ocean.

During July, the intertropical convergence zone is generally found north of the equator (**Figure 7.32B**). This shift in position occurs because the altitude of the Sun is now higher in the Northern Hemisphere. The greatest spatial shift in the ITCZ, from January to July, occurs in the eastern half of the image. This shift is about 40° of latitude in some places. The more intense July Sun causes land areas of Northern Africa and Asia rapidly warm creating the Asiatic Low, which becomes part of the ITCZ. In the winter months, the intertropical convergence zone is pushed south by the development of an intense high-pressure system over central Asia (compare **Figures 7.32A** and **7.32B**). The extreme movement of the ITCZ in this part of the world also helps to intensify the development of a regional winds system called the Asian monsoon.

In reality, the subtropical high-pressure zone does not form a uniform area of high pressure stretching around the world. Instead, the system consists of several localized anticyclonic cells of high pressure. These systems are located roughly at about 20 to 30° of latitude and are labeled with the letter H. The subtropical high-pressure systems develop because of the presence descending air currents from the Hadley cell. These systems intensify over the ocean during the summer or high Sun season. During this season, the air over the ocean bodies remains relatively cool because of the slower heating of water relative to land surfaces. Over land, intensification takes place in the winter months. At this time, land cools off quickly, relative to ocean, forming large cold continental air masses.

The subpolar lows form a continuous zone of low pressure at latitudes between 50 and 70° in the Southern Hemisphere. The location and intensity of the subpolar lows varies with season. This zone is most intense during Southern Hemisphere summer (**Figure 7.32A**). At this time, greater differences in temperature exist between air masses found on either side of this zone. North of the subpolar low belt, summer heating warms subtropical air masses. South of the zone, the ice-covered surface of Antarctica reflects much of the incoming solar radiation back to space. As a consequence, air masses above Antarctica remain cold because very little heating of the ground surface takes place. The meeting of the warm subtropical and cold polar air masses at the subpolar low zone enhances frontal uplift and the formation of intense

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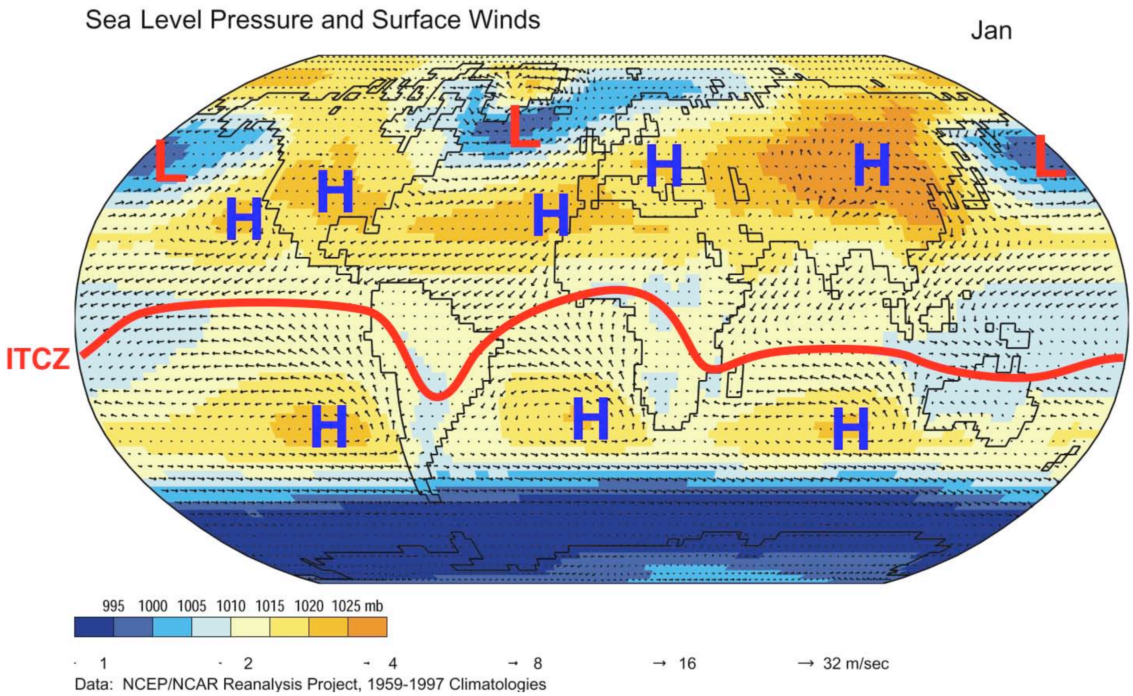
**A. January**




**B. July**

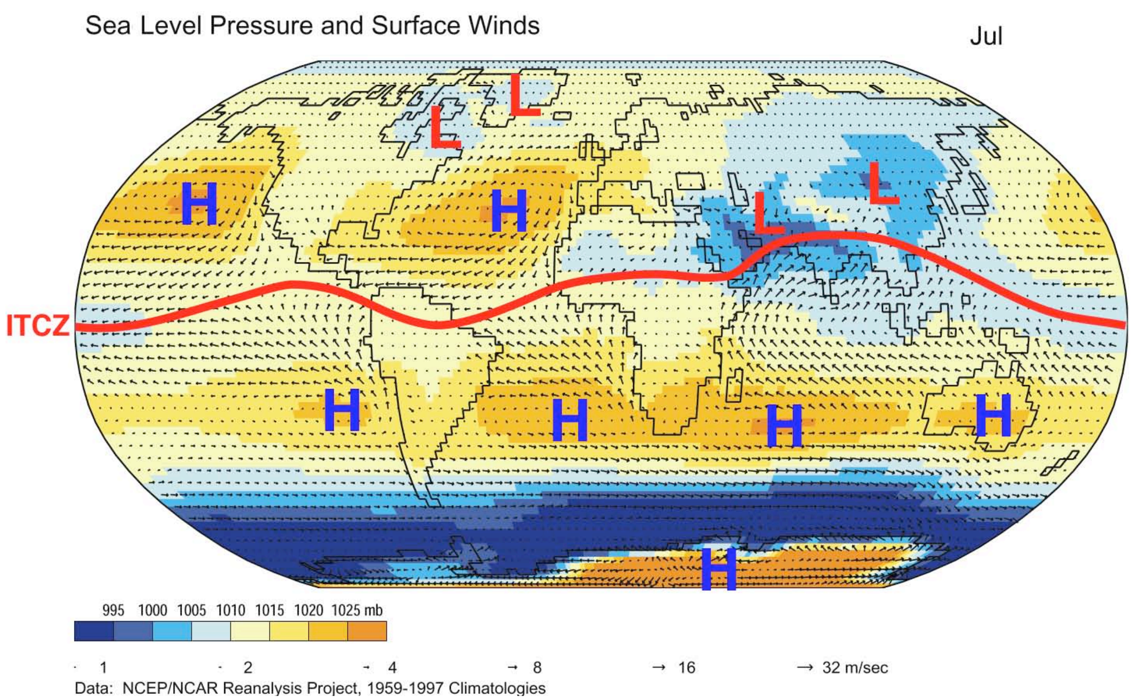



**FIGURE 7.32** January and July average surface wind patterns on the Earth's surface, 1959-1997. Longer arrows indicate faster wind speeds – see legend above. *(Original figure courtesy of J.J. Shinker, Department of Geography, University of Oregon)*

low-pressure systems.

In the Northern Hemisphere, the subpolar lows do not form a continuous belt circling the globe. Instead, they exist as localized cyclonic centers of low pressure. In the Northern Hemisphere winter, these pressure centers are intense and located over the oceans just to the south of Greenland and the Aleutian Islands (**Figure 7.32A**). These areas of low pressure are responsible for spawning many mid-latitude cyclones. The development of the subpolar lows in summer only occurs weakly (**Figure 7.32B** - over Greenland and Baffin Island, Canada), unlike the Southern Hemisphere. The reason for this phenomenon is that considerable heating of the Earth's surface occurs from 60 to 90° North. As a result, cold polar air masses generally do not form in the summer.

## UPPER ATMOSPHERE CIRCULATION AND JET STREAMS

Horizontal patterns of atmospheric circulation above the Earth's surface vary with elevation. In the troposphere, upper atmosphere flow is poleward with a predominant east to west direction. Because of the directional consistency of these winds, they are commonly called the **upper air westerlies**. The upper air westerlies form because air pressure is generally high over equatorial regions and low over the Earth's polar regions. Consequently, this spatial configuration causes the development of a global scale pressure gradient. Pressure gradient force then moves air aloft from the equator to the poles. The strong east to west component of this airflow is caused by the deflection imparted on the moving air by the Coriolis effect. Because friction is at a minimum in the upper atmosphere, the deflection is at a maximum causing the winds in the upper atmosphere to flow roughly parallel to the isobars.

The upper air westerlies are quite dominant from 30 to 80° North and South of the equator (**Figure 7.33**). Within this region of the atmosphere there are also two narrow zones in each hemisphere where the upper air westerlies become extremely fast. These zones are known as **jet streams**. In these narrow regions of the upper air westerlies, winds speeds accelerate because of the local presence of steep pressure gradients. **Figure 7.34** shows the relative position of the jet streams in each hemisphere. Note that these meandering bands of fast moving air are unbroken, circling the entire planet.

The best known of the jet streams is called the **polar jet stream**. This jet stream was discovered around the time of World War II when airplanes began flying at higher altitudes. American pilots flying bombing missions to Japanese-occupied Pacific islands occasionally encountered these winds on their flights. This encounter would often cause the planes to make little headway when they were flying westward (into the prevailing jet stream winds). On return flights (eastward in direction) back to the U.S. bases flight speeds would be accelerated because of strong tail winds produced by the jet stream. The polar jet stream occurs in the middle latitudes at an altitude of between 8 and 12 km (5 and 7.5 mi). The width of the polar jet stream varies between 100 to 500 km (60 to 300 mi). Vertical thickness is usually between 1000 to 2000 m (3250 to 6500 ft). Wind speeds at the core of this feature often are greater than 200 kph (120 mph). Maximum speeds can occasionally reach 400 kph (240 mph). The jet stream core is surrounded by a zone of slower moving air that has an average velocity of between 75 to 125 kph (45 to 78 mph).

On an upper-air weather chart, the polar jet stream resembles a meandering stream transporting large quantities of air from west to east. Analyzing upper-air weather charts over a one-year period reveals that the polar jet stream experiences considerable daily and seasonal variations in its position. In the winter months, the polar stream shifts its average position in a southerly direction. At the peak of winter, the polar jet stream can occasionally influence locations with a latitude of 30° (for example, places like northern Mexico and Florida). The polar jet stream migrates towards the poles during the spring and

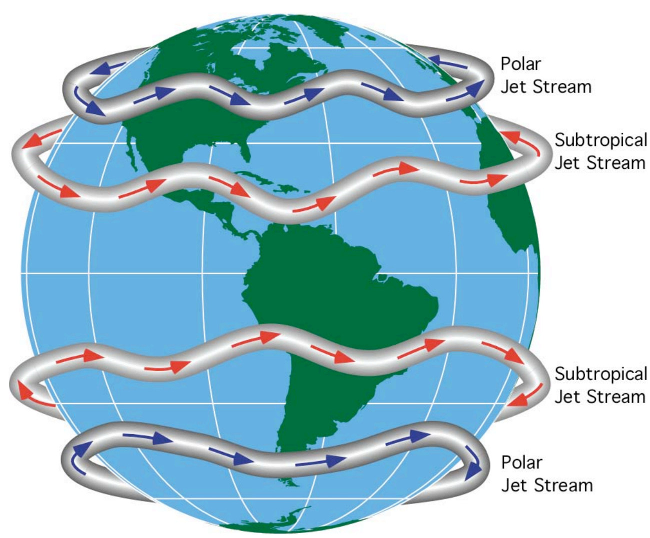
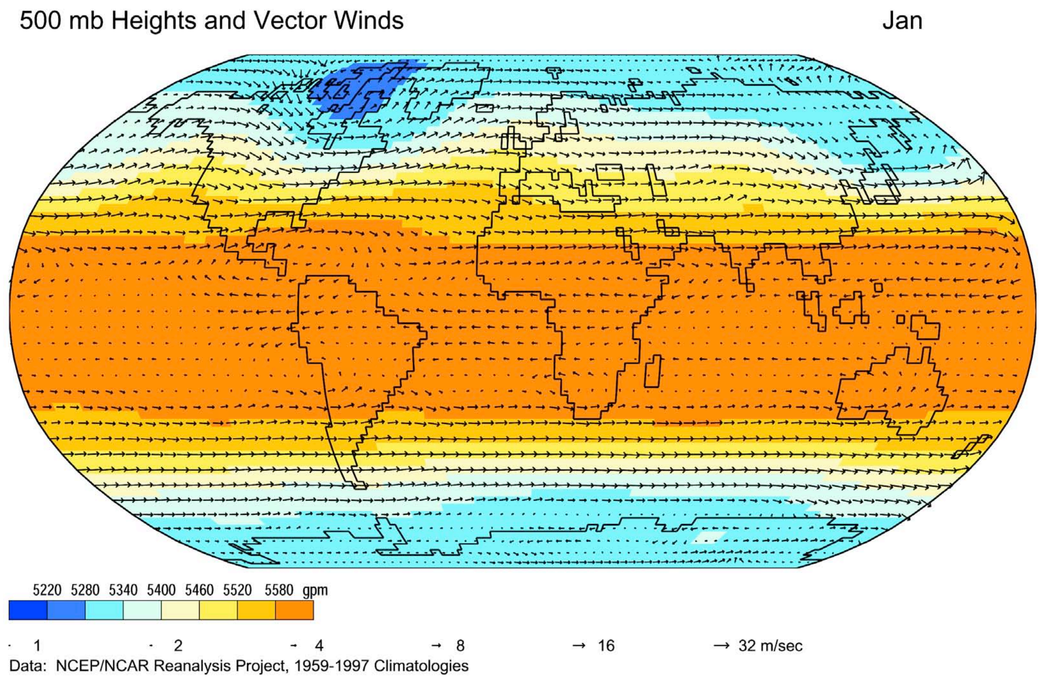



**FIGURE 7.34** Polar and subtropical jet streams. *(Image Copyright: Michael Pidwirny)*

## A. January




## B. July

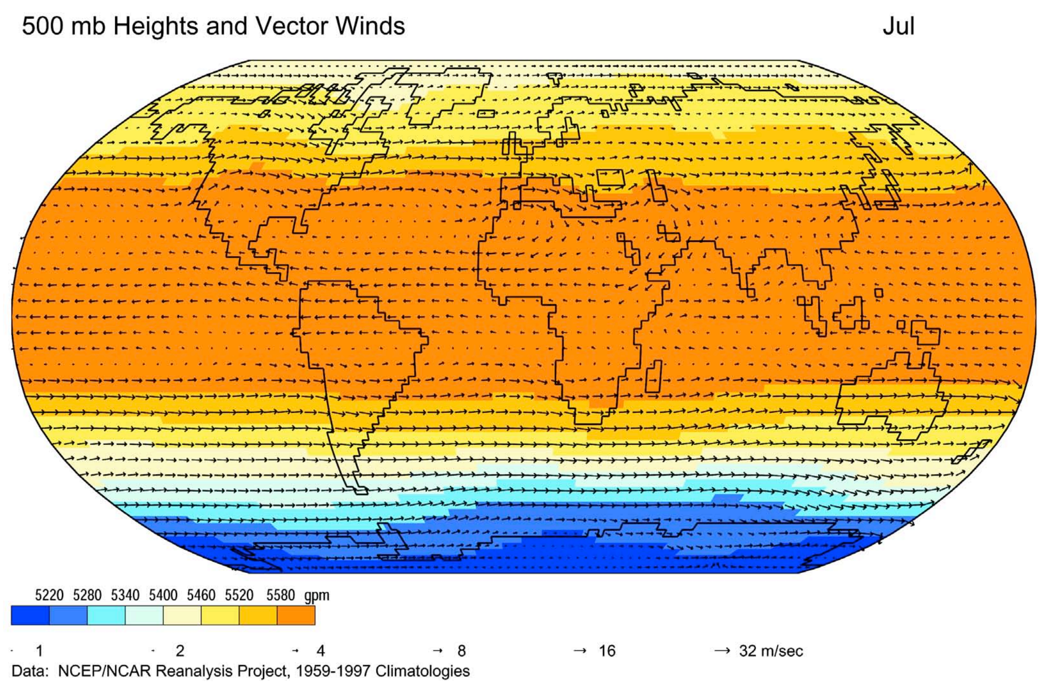



**FIGURE 7.33** January and July average wind speed and direction at the 500 millibar level, 1959-1997. The 500 millibar level represents an altitude of about 5400 meters (17,700 feet). Longer arrows indicate faster winds – see legend above. *(Original figure courtesy of J.J. Shinker, Department of Geography, University of Oregon)*

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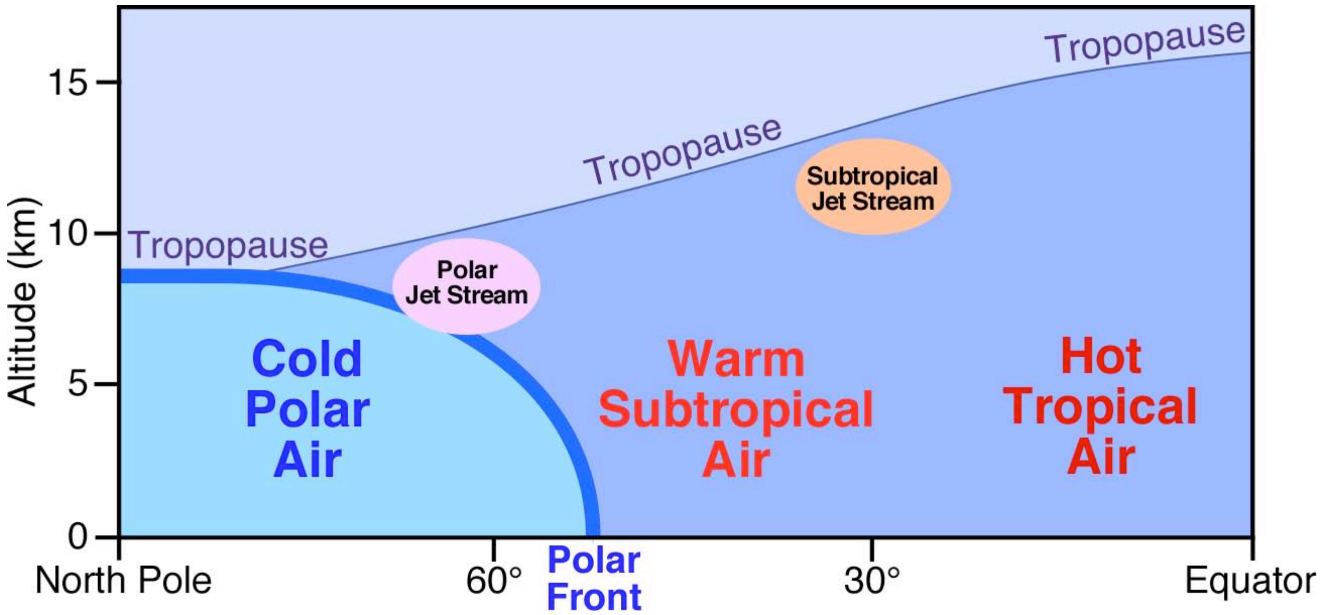


**FIGURE 7.35** Vertical cross-section of the atmosphere from the North Pole to the equator showing the relative location of the polar front and the polar jet stream. The polar front is a transition zone where cold polar air and warm subtropical air meet. *(Image Copyright: Michael Pidwirny)*

summer months. The average position of the polar jet stream during summer months is about 55° latitude (north of the Canada/USA border).

Associated with the polar jet stream is the polar front. The **polar front** represents the transitional zone where warm air from the subtropics and cold air from the poles meet (**Figure 7.35**). At this zone, massive exchanges of energy occur in the form of cyclonic storms known as the **mid-latitude cyclones**. The shape and position of waves in the polar jet stream determine the location and the intensity of the mid-latitude cyclones. In general, mid-latitude cyclones form beneath polar jet stream troughs. The following satellite image (**Figure 7.36**), taken from above the South Pole, shows a number of mid-latitude cyclones circling Antarctica. Each mid-latitude cyclone wave is defined by the cloud development associated with frontal uplift.

The **subtropical jet stream** lies approximately 13 km (8 mi) above the subtropical high-pressure zone. The reason for its formation is similar to the polar jet stream: it develops because of the presence of a localized steep pressure gradient in the upper atmosphere. The subtropical jet stream differs from the polar jet stream in two distinct ways. First, the subtropical jet stream has generally lower wind speeds. During summer season, the velocity of the subtropical jet stream can drop to a point where the feature blends in with the surrounding upper air westerlies. This condition can last for several days. Secondly, the subtropical jet stream tends to show less day-to-day and seasonal variations in latitudinal position. Sometimes in the winter and early spring, the subtropical jet stream can join with the polar jet stream. These mergers are mainly due to

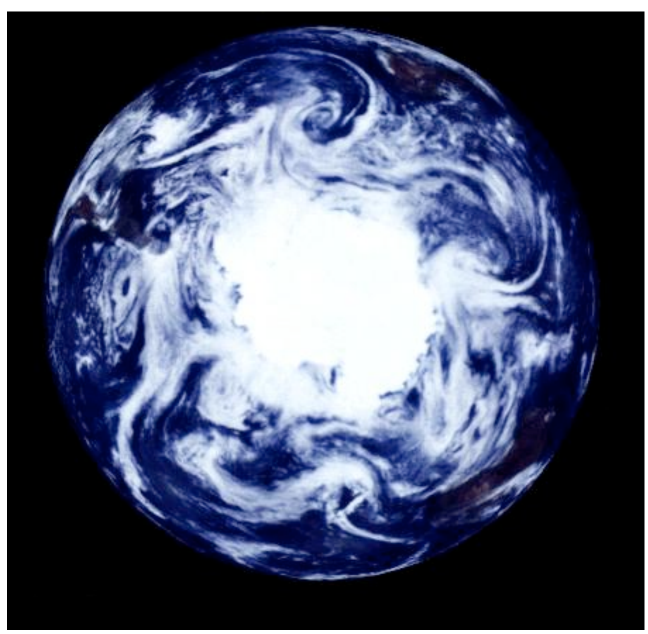

**FIGURE 7.36** Satellite view of the atmospheric circulation at the South Pole. The spiraling waves of clouds circling around Antarctica are mid-latitude cyclones. *(Source: NASA)*

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the migration of the polar jet stream into areas where the subtropical jet stream normally exists.

## Rossby Waves

It is important to recognize that the polar jet stream is a connected part of a much bigger airflow: the upper air westerlies. Because of this connection, undulations in the polar jet stream indicate much larger waves in global atmospheric circulation. Another name for these large waves in the upper atmosphere is **Rossby waves**, named after meteorologist C.G. Rossby who studied this atmospheric feature in the 1930s.

Most of our knowledge about the nature of Rossby waves comes from the research work of C.G. Rossby. Normally, there are between three to seven Rossby waves circling the planet at any time. The characteristics of individual waves in this atmospheric system can vary considerably in terms of wavelength (the distance separating adjacent wave ridges or troughs), amplitude (latitudinal extent), and internal wind speeds (**Figure 7.37**). Variations in Rossby wave number tend to be related to seasonal variations in climate. During the winter season, latitudinal extremes in temperature generally create fewer waves with greater amplitudes and stronger associated winds. Summer is associated with more waves with small amplitudes and weak upper air winds.

Rossby waves can have a great influence on the daily weather of locations in the middle and high latitudes. The development of waves in this system indicates the movement of warm subtropical air poleward and cold polar air towards equator. This redistribution of cold and warm air can cause drastic changes in weather when waves move into an area.

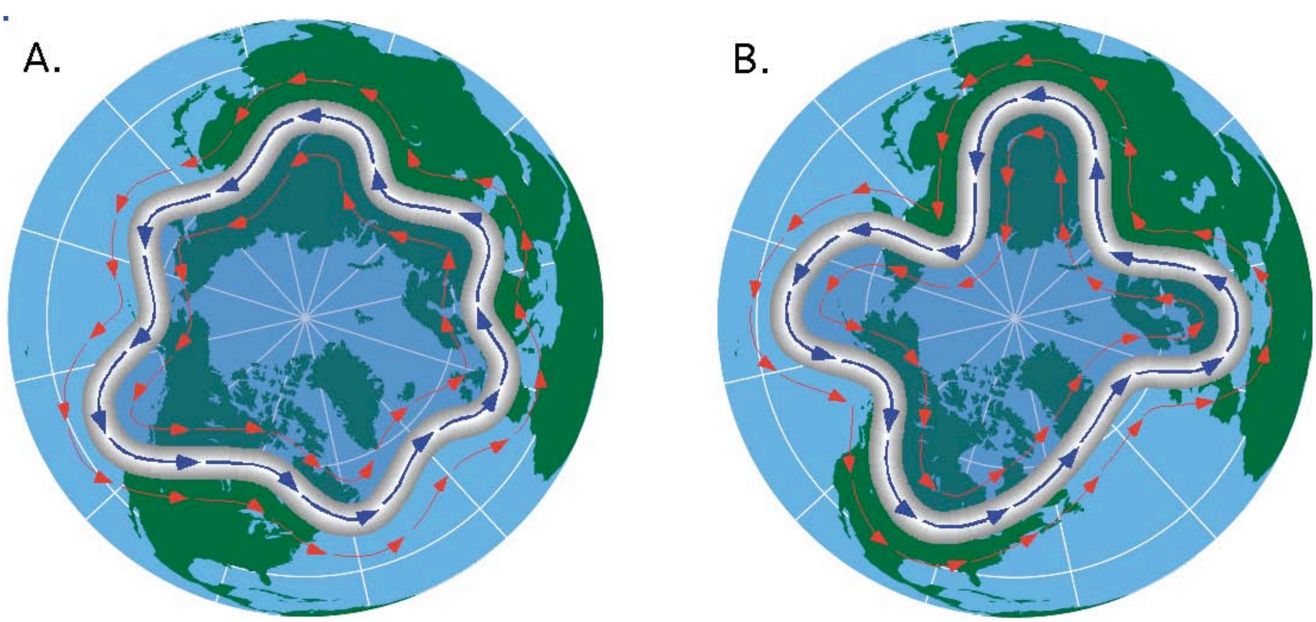


**FIGURE 7.37** Rossby waves in the upper atmosphere as defined by the polar jet stream and associated upper air circulation. In illustration A, six waves exist that have relatively small amplitudes.. Illustration B has only four waves that have much larger amplitudes. Because of the greater amplitudes of the much larger waves, excursions of warm air towards the poles and cold air towards the equator travel greater distances. *(Image Copyright: Michael Pidwirny)*

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## CHAPTER SUMMARY

- Atmospheric pressure can be simply defined as the weight of the atmosphere. Measurements of atmospheric pressure are made regularly by meteorologists to create weather maps used to forecasts future weather conditions.
- Any instrument used to measure atmospheric pressure is called a barometer. Barometers measure air pressure in a variety of units including inches, millibars, pascals, and kilopascals.
- Measurements of air pressure vary over space and time due to a variety of factors. Vertical variations in the atmosphere are mainly due to changes in the quantity of air lying above the location being measured.
- The heating and cooling of air, changes in water vapor concentrations, and atmospheric circulation are the primary factors responsible for changes in air pressure near the Earth's surface.
- Weather maps use a system of lines called isobars to illustrate variations in air pressure near the Earth's surface and within the upper atmosphere.
- On global scale maps of surface atmospheric pressure, one can observe seasonal patterns of low and high pressure cells that return to the same regions of the planet year-after-year.
- Wind is the movement of air in the Earth's atmosphere. This movement develops because of spatial differences in atmospheric pressure. Newton's laws of motion suggest that wind should blow from areas of high density to areas of low density.
- The speed of wind flow is controlled by pressure gradient force. Pressure gradient force can be simply described as the rate of pressure change (pressure gradient) over space.
- Coriolis effect refers to an apparent force that causes medium and large-scale movements of air to be deflected from their intended path. This deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere.
- Latitude of the location and the speed of the wind determine the strength of the Coriolis effect.
- Frictional force influences moving air by reducing its speed. However, friction only acts on wind that is flowing near the surface of the Earth.
- Atmospheric scientists have described a variety of different types of large-scale winds. A geostrophic wind occurs in regions of the atmosphere (1 km above the Earth's surface) where friction is low and the air tends to flow in a straight path. In geostrophic winds, only two forces are active: pressure gradient force and Coriolis effect.
- Winds blowing in the upper atmosphere in curved paths are called gradient winds. Winds near the Earth's surface are called friction layer winds. Friction layer winds are the result of pressure gradient force, Coriolis effect, and frictional force.
- There are a number of unique types of circulation that exist at local and regional scales. These winds are often the result of thermally generated circulation systems. In these systems, warm temperatures create areas of low atmospheric pressure on the Earth's surface, while areas of high pressure are generated when temperatures are relatively cold. Once established, the winds associated with a thermal gradient move from high to low pressure on the Earth's surface.
- Winds that are the result of thermal gradients include, land and sea breeze, mountain and valley breeze, and on a regional scale monsoon winds.
- Katabatic winds are a special type of mountain breeze that involve the relatively fast flow of cold air downslope.
- At the global scale, consistent patterns of airflow can be seen at the Earth's surface and within its upper atmosphere. Global scale patterns of circulation develop because of latitudinal variations in atmospheric pressure. However, these differences in pressure are not just the result of the differential heating of the Earth's surface.
- The intertropical convergence zone (ITCZ) is normally found near the equator. The ITCZ is created by combined effects of trade wind convergence and thermal convection.
- At approximately 30° North and South latitude, the subtropical high-pressure zone forms because of the presence of descending air from the upper atmosphere.
- The subpolar lows, located at about 60° North and South latitude, develop because of the dynamic interaction of cold polar air with warm moist subtropical air masses.
- Surface winds move from areas high pressure to low pressure. This course of movement is also altered by the influence of Coriolis effect causing the development of the trade winds (0 to 30°N and S), the westerlies (30 to 60°N and S) and the polar easterlies (60 to 90°N and S).
- Upper atmosphere winds are generally poleward and westerly direction. Their development is related to the presence of the Hadley, Ferrel, and Polar circulation cells in the North and South Hemisphere.
- Associated with the upper air westerlies are narrow regions of intensification where fast moving air is channeled into west to east flowing jet streams. Two different jet streams exist in each hemisphere.

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- Over the subtropical high zones is the subtropical jet stream. The more intense polar jet stream has a position roughly located above the subpolar lows.
- Associated with the upper air westerlies and the polar jet stream are the Rossby waves. Rossby waves control the movement of large masses of cold and warm air in the middle and high latitudes.

## IMPORTANT TERMS

**Anemometer**
**Aneroid barometer**
**Anticyclones**
**Atmospheric pressure**
**Barograph**
**Barogram**
**Barometer**
**Convection**
**Convergence**
**Coriolis effect**
**Cyclones**
**Eddies**
**Ferrel Cell**
**First law of motion**
**Friction layer**
**Frictional force**
**Geostrophic wind**
**Gradient wind**
**Gravity**
**Hadley Cell**
**Ideal Gas Law**
**Intertropical convergence zone (ITCZ)**
**Isobar**
**Katabatic wind**
**KiloPascal (kPa)**
**Land breeze**
**Mass**
**Meridional**
**Mid-latitude cyclone**
**Millibar (mb)**
**Monsoon**
**Mountain breeze**
**Northeast Trades**
**Orographic uplift**
**Pascal (Pa)**
**Polar Cell**
**Polar front**
**Polar high**
**Polar jet stream**
**Pressure gradient force**
**Radiosonde**
**Rossby waves**
**Sea breeze**
**Second law of motion**
**Southeast Trades**
**Subpolar lows**
**Subtropical high-pressure zone**
**Subtropical jet stream**
**Upper air westerlies**
**Valley breeze**
**Venturi effect**
**Weight**
**Westerlies**
**Wind**
**Wind direction**
**Wind speed**
**Wind vane**

## CHAPTER REVIEW QUESTIONS

1. Explain the concept of atmospheric pressure. Why does air pressure vary with altitude? What factors cause surface air pressure to fluctuate across time and space?
2. What is the Ideal Gas Law? According to this law, describe the effect changes to air temperature, volume, and density have on air pressure.
3. What instruments and units are used to measure air pressure? How are surface air pressure readings adjusted to make them more comparable?
4. How are surface atmospheric pressure measurements depicted on weather maps? On a map of the world, describe the patterns of monthly average sea level pressure that occur on the Earth's surface for January and July.

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5. What role does pressure gradient force have the formation of wind? How does Coriolis effect and frictional force influence wind near the ground surface and in the upper atmosphere?
6. Describe how localized differences in surface heating can result in the development in a thermal circulation cell.
7. Why is land/sea breeze considered a thermal circulation system? Explain how land breeze differs from sea breeze.
8. How do mountain and valley breezes form? What is a katabatic wind? Why is this type of circulation system also known as a drainage wind?
9. What is a monsoon? Describe the factors responsible for the formation of summer and winter monsoon winds in Southeast Asia.
10. With the aid of a diagram, explain the "Three Cell Model of Global Circulation". On this diagram indicate and name zones of dominant surface high and low pressure cells, prevailing surface wind systems, and the location and flow characteristics of the Hadley, Ferrel, and Polar cell. How does this model differ from the Earth's actual global surface circulation?
11. Describe the characteristics of the following upper atmosphere circulation features: upper air westerlies, polar jet stream, and subtropical jet stream.
12. What is a Rossby wave? How does this feature influence surface weather in the middle and high latitudes?

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