CHAPTER 8

Air Pressure and Winds

December 19, 1980, was a cool day in Lynn, Massachusetts, but not cool enough to dampen the spirits of more than 2000 people who gathered in Central Square — all hoping to catch at least one of the 1500 dollar bills that would be dropped from a small airplane at noon. Right on schedule, the aircraft circled the city and dumped the money onto the people below. However, to the dismay of the onlookers, a westerly wind caught the currency before it reached the ground and carried it out over the cold Atlantic Ocean. Had the pilot or the sponsoring leather manufacturer examined the weather charts beforehand, it might have been possible to predict that the wind would ruin the advertising scheme.

CONTENTS

This opening scenario raises two questions: (1) Why does the wind blow? and (2) How can one tell its direction by looking at weather charts? Chapter 1 has already answered the first question: Air moves in response to horizontal differences in pressure. This phenomenon happens when we open a vacuum-packed can—air rushes from the higher pressure region outside the can toward the region of lower pressure inside. In the atmosphere, the wind blows in an attempt to equalize imbalances in air pressure. Does this mean that the wind always blows directly from high to low pressure? Not really, because the movement of air is controlled not only by pressure differences but by other forces as well.

In this chapter, we will first consider how and why atmospheric pressure varies, then we will look at the forces that influence atmospheric motions aloft and at the surface. Through studying these forces, we will be able to tell how the wind should blow in a particular region by examining surface and upper-air charts.

Atmospheric Pressure

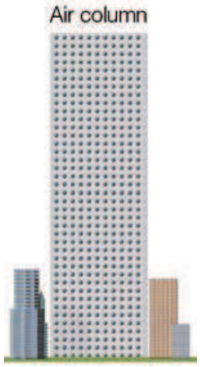
In Chapter 1, we learned several important concepts about atmospheric pressure. One stated that **air pressure** is simply the mass of air above a given level. As we climb in elevation above the earth's surface, there are fewer air molecules above us; hence, atmospheric pressure always decreases with increasing height. Another concept we learned was that most of our atmosphere is crowded close to the earth's surface, which causes air pressure to decrease with height, rapidly at first, then more slowly at higher altitudes.

So one way to change air pressure is to simply move up or down in the atmosphere. But what causes the air pressure to change in the horizontal? And why does the air pressure change at the surface?

HORIZONTAL PRESSURE VARIATIONS—A TALE OF TWO CITIES

To answer these questions, we eliminate some of the complexities of the atmosphere by constructing *models.* ● Figure 8.1 shows a simple atmospheric model—a column of air, extending well up into the atmosphere. In the column, the dots represent air molecules. Our model assumes: (1) that the air molecules are not crowded close to the surface and, unlike the real atmosphere, the air density remains constant from the surface up to the top of the column, (2) that the width of the column does not change with height and (3) that the air is unable to freely move into or out of the column.

Suppose we somehow force more air into the column in Fig. 8.1. What would happen? If the air temperature in the column does not change, the added air would make the column more dense, and the added weight of the air in the column would increase the surface air pressure. Likewise, if a great deal of air were removed from the column, the surface air pressure would decrease. Consequently, to change the surface air pressure, we need to change the mass of air in the column above the surface. But how can this feat be accomplished?



● FIGURE 8.1 A model of the atmosphere where air density remains constant with height. The air pressure at the surface is related to the number of molecules above. When air of the same temperature is stuffed into the column, the surface air pressure rises. When air is removed from the column, the surface pressure falls.

Look at the air columns in ● Fig. 8.2a.* Suppose both columns are located at the same elevation, both have the same air temperature, and both have the same surface air pressure. This condition, of course, means that there must be the same number of molecules (same mass of air) in each column above both cities. Further suppose that the surface air pressure for both cities remains the same, while the air above city 1 cools and the air above city 2 warms (see Fig. 8.2b).

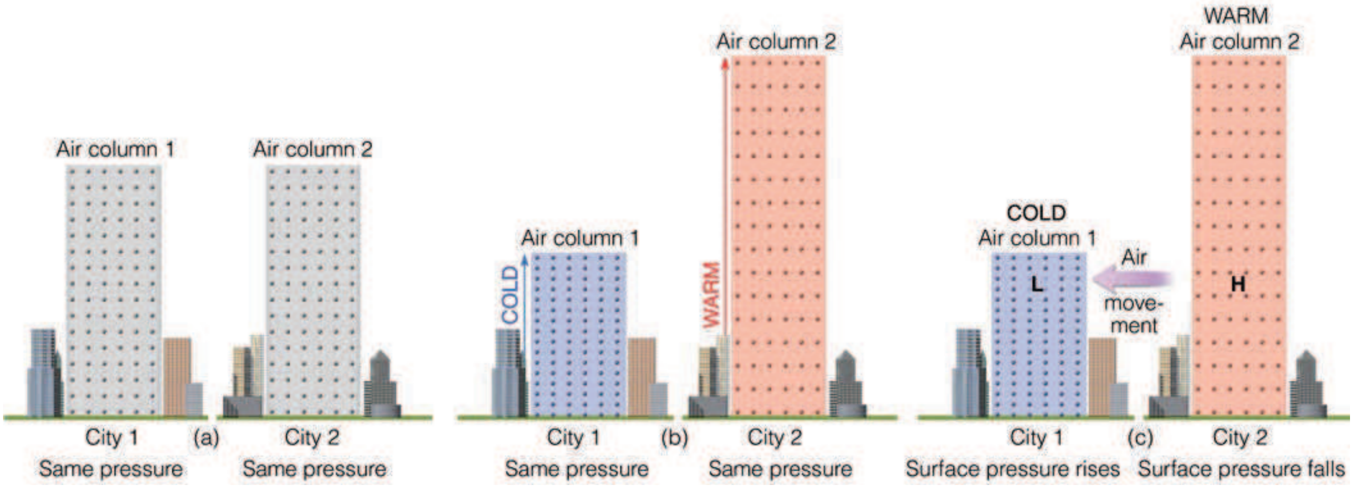
As the air in column 1 cools, the molecules move more slowly and crowd closer together—the air becomes more dense. In the warm air above city 2, the molecules move faster and spread farther apart—the air becomes less dense. Since the width of the columns does not change (and if we assume an invisible barrier exists between the columns), the total number of molecules above each city remains the same, and the surface pressure does not change. Therefore, in the more-dense cold air above city 1, the column shrinks, while the column rises in the less-dense, warm air above city 2.

We now have a cold, shorter dense column of air above city 1 and a warm, taller less-dense air column above city 2. From this situation, we can conclude that *it takes a shorter column of cold, more-dense air to exert the same surface pressure as a taller column of warm, less-dense air.* This concept has a great deal of meteorological significance.

Atmospheric pressure decreases more rapidly with height in the cold column of air. In the cold air above city 1 (Fig. 8.2b), move up the column and observe how quickly you pass through the densely packed molecules. This activity indicates a rapid change in pressure. In the warmer, less-dense air, the pressure does not decrease as rapidly with height, simply because you climb above fewer molecules in the same vertical distance.

In Fig. 8.2c, move up the warm, red column until you come to the letter *H*. Now move up the cold, blue column the same distance until you reach the letter *L*. Notice that there are more molecules above the letter *H* in the warm column than above the letter *L* in the cold column. The fact that the

*We will keep our same assumption as in Fig. 8.1; that is, (1) the air molecules are not crowded close to the surface, (2) the width of the columns does not change, and (3) air is unable to move into or out of the columns.



ACTIVE FIGURE 8.2 (a) Two air columns, each with identical mass, have the same surface air pressure. (b) Because it takes a shorter column of cold air to exert the same pressure as a taller column of warm air, as column 1 cools, it must shrink, and as column 2 warms, it must expand. (c) Because at the same level in the atmosphere there is more air above the H in the warm column than above the L in the cold column, warm air aloft is associated with high pressure and cold air aloft with low pressure. The pressure differences aloft create a force that causes the air to move from a region of higher pressure toward a region of lower pressure. The removal of air from column 2 causes its surface pressure to drop, whereas the addition of air into column 1 causes its surface pressure to rise. (The difference in height between the two columns is greatly exaggerated.) Visit the Meteorology Resource Center to view this and other active figures at academic.cengage.com/login

number of molecules above any level is a measure of the atmospheric pressure leads to an important concept: *Warm air aloft is normally associated with high atmospheric pressure, and cold air aloft is associated with low atmospheric pressure.*

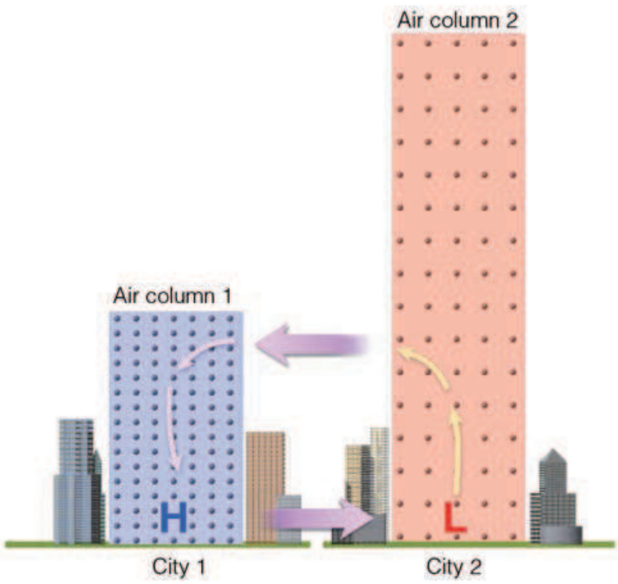
In Fig. 8.2c, the horizontal difference in temperature creates a horizontal difference in pressure. The pressure difference establishes a force (called the *pressure gradient force*) that causes the air to move from higher pressure toward lower pressure. Consequently, if we remove the invisible barrier between the two columns and allow the air aloft to move horizontally, the air will move from column 2 toward column 1. As the air aloft leaves column 2, the mass of the air in the column decreases, and so does the surface air pressure. Meanwhile, the accumulation of air in column 1 causes the surface air pressure to increase.

Higher air pressure at the surface in column 1 and lower air pressure at the surface in column 2 causes the surface air to move from city 1 towards city 2 (see ● Fig. 8.3). As the surface air moves out away from city 1, the air aloft slowly sinks to replace this outwardly spreading surface air. As the surface air flows into city 2, it slowly rises to replace the depleted air aloft. In this manner, a complete circulation of air is established due to the heating and cooling of air columns.

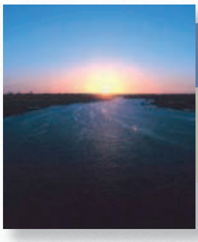
In summary, we can see how heating and cooling columns of air can establish horizontal variations in air pressure both aloft and at the surface. It is these horizontal differences in air pressure that cause the wind to blow.

Air temperature, air pressure, and air density are all interrelated. If one of these variables changes, the other two usually change as well. The relationship among these three variables is expressed by the gas law, which is described in the Focus section on p. 196.

DAILY PRESSURE VARIATIONS From what we have learned so far, we might expect to see the surface pressure dropping as the air temperature rises, and vice versa. Over large continental areas, especially the southwestern United States in summer, hot surface air is accompanied by surface low pressure. Likewise, bitter cold arctic air in winter is often accom-



● FIGURE 8.3 The heating and cooling of air columns causes horizontal pressure variations aloft and at the surface. These pressure variations force the air to move from areas of higher pressure toward areas of lower pressure. In conjunction with these horizontal air motions, the air slowly sinks above the surface high and rises above the surface low.



FOCUS ON A SPECIAL TOPIC

The Atmosphere Obeys the Gas Law

The relationship among the pressure, temperature, and density of air can be expressed by

$$\text{Pressure} = \text{temperature} \times \text{density} \times \text{constant}.$$

This simple relationship, often referred to as the *gas law* (or *equation of state*), tells us that the pressure of a gas is equal to its temperature times its density times a constant. When we ignore the constant and look at the gas law in symbolic form, it becomes

$$p \sim T \times \rho,$$

where, of course, p is pressure, T is temperature, and ρ (the Greek letter rho, pronounced "row") represents air density.* The line ~ is a symbol meaning "is proportional to." A change in one variable causes a corresponding change in the other two variables. Thus, it will be easier to understand the behavior of a gas if we keep one variable from changing and observe the behavior of the other two.

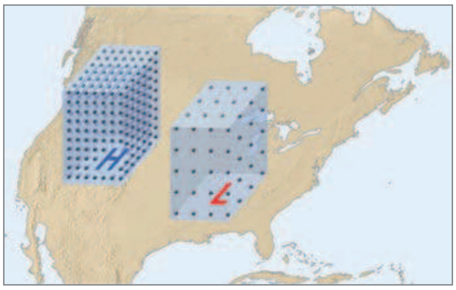
Suppose, for example, we hold the temperature constant. The relationship then becomes

$$p \sim \rho \text{ (temperature constant).}$$

This expression says that the pressure of the gas is proportional to its density, as long as its temperature does not change. Consequently, if the temperature of a gas (such as air) is held constant, as the pressure increases the density increases, and as the pressure decreases the density decreases. In other words, *at the same temperature, air at a higher pressure is more dense than air at a lower pressure.* If we apply this concept to the atmosphere, then with nearly the same temperature and elevation, air above a region of surface high pressure is more dense than air above a region of surface low pressure (see Fig. 1).

We can see, then, that for surface high-pressure areas (anticyclones) and surface low-pressure areas (mid-latitude cyclones) to form,

*This gas law may also be written as $p \times v = T \times$ constant. Consequently, pressure and temperature changes are also related to changes in volume.



● FIGURE 1 Air above a region of surface high pressure is more dense than air above a region of surface low pressure (at the same temperature). (The dots in each column represent air molecules.)

the air density (mass of air) above these systems must change. As we will see later in this chapter, as well as in other chapters, surface air pressure increases when the wind causes more air to move into a column of air than is able to leave (called *net convergence*), and surface air pressure decreases when the wind causes more air to move out of a column of air than is able to enter (called *net divergence*).

Earlier, we considered how pressure and density are related when the temperature is not changing. What happens to the gas law when the pressure of a gas remains constant? In shorthand notation, the law becomes

$$(\text{Constant pressure}) \times \text{constant} = T \times \rho.$$

This relationship tells us that when the pressure of a gas is held constant, the gas becomes less dense as the temperature goes up, and more dense as the temperature goes down. Therefore, *at a given atmospheric pressure, air that is cold is more dense than air that is warm.* Keep in mind that the idea that cold air is more dense than warm air applies only when we compare volumes of air at the same level, where pressure changes are small in any horizontal direction.

We can use the gas law to obtain information about the atmosphere. For example, at an altitude of about 5600 m (18,400 ft) above sea level, the atmospheric pressure is normally close to 500 millibars. If we obtain the average density at this level, with the aid of the gas law we can calculate the average air temperature.

Recall that the gas law is written as

$$p = T \times \rho \times C.$$

With the pressure (p) in millibars (mb), the temperature (T) in Kelvins, and the density (ρ) in kilograms per cubic meter (kg/m^3), the numerical value of the constant (C) is about 2.87.*

At an altitude of 5600 m above sea level, where the average (or standard) air pressure is about 500 mb and the average air density is 0.690 kg/m^3, the average air temperature becomes

$$p = T \times \rho \times C$$
$$500 = T \times 0.690 \times 2.87$$
$$\frac{500}{0.690 \times 2.87} = T$$
$$252.5 \text{ K} = T.$$

To convert Kelvins into degrees Celsius, we subtract 273 from the Kelvin temperature and obtain a temperature of −20.5°C, which is the same as −5°F.

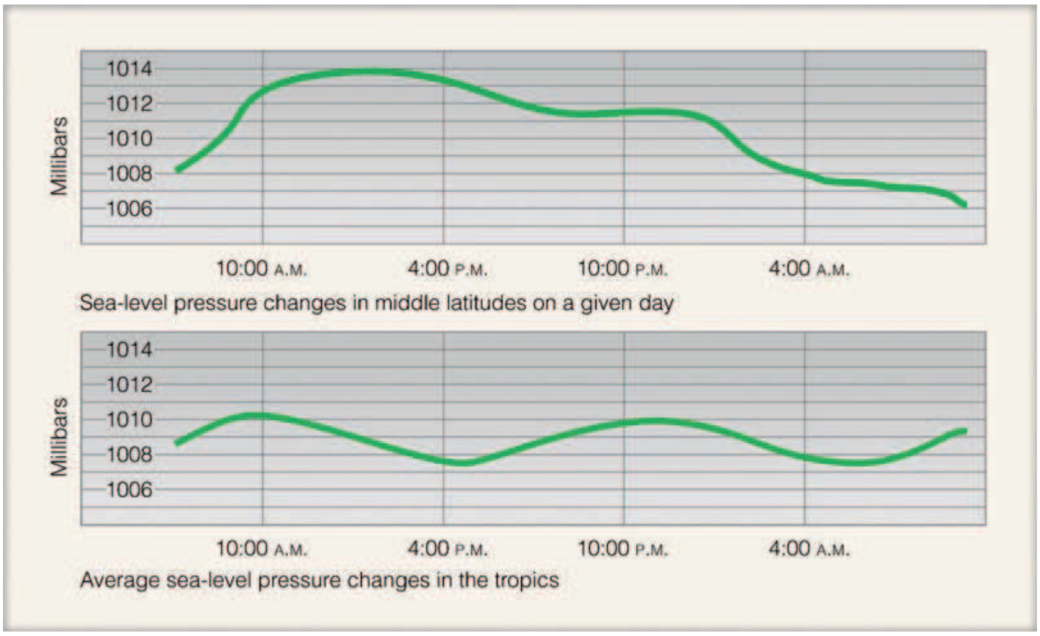
If we know the numerical values of temperature and density, with the aid of the gas law we can obtain the air pressure. For example, in Chapter 1 we saw that the average global temperature near sea level is 15°C (59°F), which is the same as 288 K. If the average air density at sea level is 1.226 kg/m^3, what would be the standard (average) sea-level pressure?

Using the gas law, we obtain

$$p = T \times \rho \times C$$
$$p = 288 \times 1.226 \times 2.87$$
$$p = 1013 \text{ mb}.$$

Since the air pressure is related to both temperature and density, a small change in either or both of these variables can bring about a change in pressure.

*The constant is usually expressed as 2.87×10^6 erg/g K, or, in the SI system, as 287 J/kg K. (See Appendix A for information regarding the units used here.)



● FIGURE 8.4 Diurnal surface pressure changes in the middle latitudes and in the tropics.

panied by surface high pressure. Yet, on a daily basis, any cyclic change in surface pressure brought on by daily temperature changes is concealed by the pressure changes created by the warming of the upper atmosphere.

In the tropics, for example, pressure rises and falls in a regular pattern twice a day (see ● Fig. 8.4). Maximum pressures occur around 10:00 A.M. and 10:00 P.M., minimum near 4:00 A.M. and 4:00 P.M. The largest pressure difference, about 2.5 mb, occurs near the equator. It also shows up in higher latitudes, but with a much smaller amplitude. This daily *(diurnal)* fluctuation of pressure appears to be due primarily to the absorption of solar energy by ozone in the upper atmosphere and by water vapor in the lower atmosphere. The warming and cooling of the air creates density oscillations known as *thermal* (or *atmospheric*) *tides* that show up as small pressure changes near the earth's surface.

In middle latitudes, surface pressure changes are primarily the result of large high- and low-pressure areas that move toward or away from a region. Generally, when an area of high pressure approaches a city, surface pressure usually rises. When it moves away, pressure usually falls. Likewise, an approaching low causes the air pressure to fall, and one moving away causes surface pressure to rise.

PRESSURE MEASUREMENTS Instruments that detect and measure pressure changes are called **barometers,** which literally means an instrument that measures bars. You may recall from Chapter 1 that a *bar* is a unit of pressure that describes a force over a given area.* Because the bar is a relatively large unit, and because surface pressure changes are normally small, the unit of pressure commonly found on surface weather maps is, as we saw in Chapter 1, the **millibar** (mb), where 1 mb = 1/1000 bar or

$$1 \text{ bar} = 1000 \text{ mb}.$$

A common pressure unit used in aviation is *inches of mercury* (Hg). At sea level, *standard atmospheric pressure** is

$$1013.25 \text{ mb} = 29.92 \text{ in. Hg} = 76 \text{ cm}.$$

As a reference, ● Fig. 8.5 compares pressure readings in inches of mercury and millibars.

The unit of pressure designated by the International System (SI) of measurement is the *pascal,* named in honor of Blaise Pascal (1632–1662), whose experiments on atmospheric pressure greatly increased our knowledge of the atmosphere. A pascal (Pa) is the force of 1 newton acting on a surface area of 1 square meter. Thus, 100 pascals equals 1 millibar. The scientific community often uses the *kilopascal* (kPa) as the unit of pressure, where 1 kPa = 10 mb. However, a more convenient unit is the **hectopascal** (hPa), as

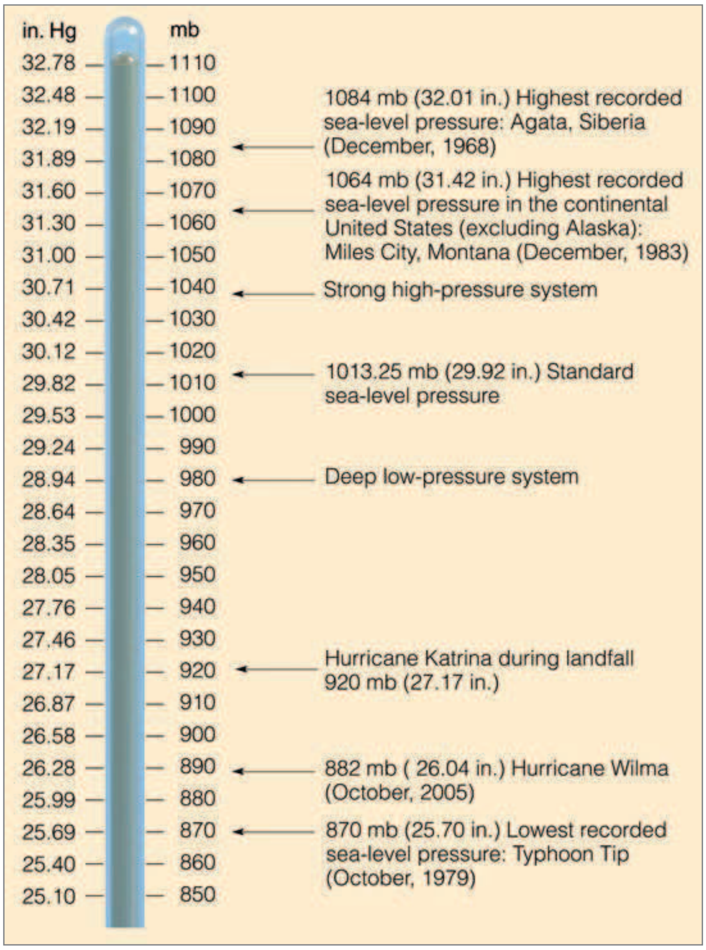
$$1 \text{ hPa} = 1 \text{ mb}.$$

WEATHER WATCH

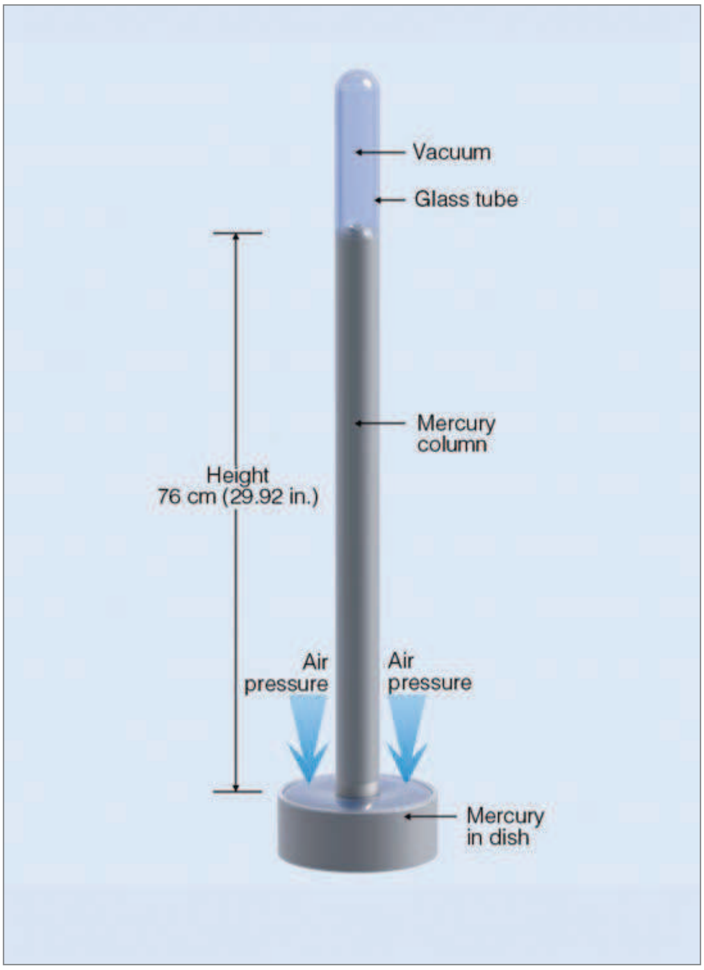
Although 1013.25 mb (29.92 in.) is the *standard atmospheric pressure* at sea level, it is *not* the average sea-level pressure. The earth's average sea-level pressure is 1011.0 mb (29.85 in.). Because much of the earth's surface is above sea level, the earth's annual average *surface pressure* is estimated to be 984.43 mb (29.07 in.).

*A bar is a force of 100,000 newtons acting on a surface area of 1 square meter. A *newton* (N) is the amount of force required to move an object with a mass of 1 kilogram so that it increases its speed at a rate of 1 meter per second each second. Additional pressure units and conversions are given in Appendix A.

*Standard atmospheric pressure at sea level is the pressure extended by a column of mercury 29.92 in. (760 mm) high, having a density of 1.36×10^4 kg/m^3, and subject to an acceleration of gravity of 9.80 m/sec^2.



● FIGURE 8.5 Atmospheric pressure in inches of mercury and in millibars.



● FIGURE 8.6 The mercury barometer. The height of the mercury column is a measure of atmospheric pressure.

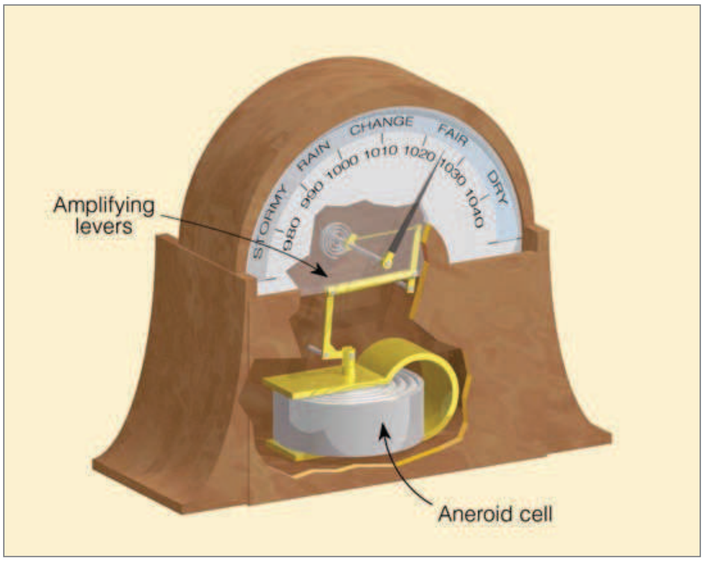
Presently, the hectopascal is gradually replacing the millibar as the preferred unit of pressure on surface weather maps.

Because we measure atmospheric pressure with an instrument called a *barometer,* atmospheric pressure is also referred to as *barometric pressure.* Evangelista Torricelli, a student of Galileo, invented the **mercury barometer** in 1643. His barometer, similar to those in use today, consisted of a long glass tube open at one end and closed at the other (see ● Fig. 8.6). Removing air from the tube and covering the open end, Torricelli immersed the lower portion into a dish of mercury. He removed the cover, and the mercury rose up the tube to nearly 76 cm (or about 30 in.) above the level in the dish. Torricelli correctly concluded that the column of mercury in the tube was balancing the weight of the air above the dish, and hence its height was a measure of atmospheric pressure.

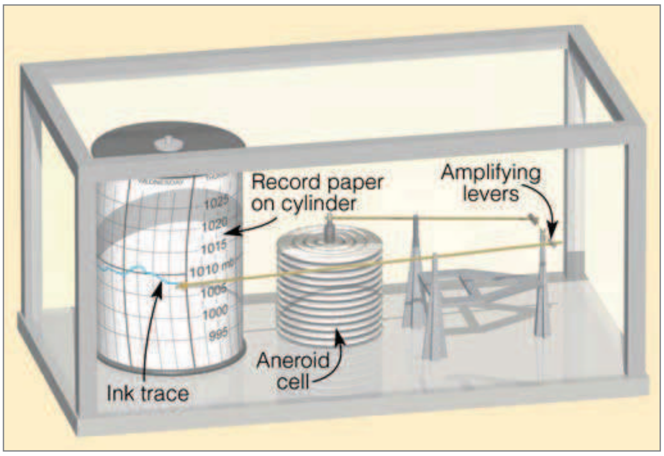
Why is mercury rather than water used in the barometer? The primary reason is convenience. (Also, water can evaporate in the tube.) Mercury seldom rises to a height above 80 cm (31.5 in.). A water barometer, however, presents a problem. Because water is 13.6 times less dense than mercury, an atmospheric pressure of 76 cm (30 in.) of mercury would be equivalent to 1034 cm (408 in.) of water. A water barometer resting on the ground near sea level would have to be read from a ladder over 10 m (33 ft) tall.

The most common type of home barometer—the **aneroid barometer**—contains no fluid. Inside this instrument is a small, flexible metal box called an *aneroid cell.* Before the cell is tightly sealed, air is partially removed, so that small changes in external air pressure cause the cell to expand or contract. The size of the cell is calibrated to represent different pressures, and any change in its size is amplified by levers and transmitted to an indicating arm, which points to the current atmospheric pressure (see ● Fig. 8.7).

Notice that the aneroid barometer often has descriptive weather-related words printed above specific pressure values. These descriptions indicate the most likely weather conditions when the needle is pointing to that particular pressure reading. Generally, the higher the reading, the more likely clear weather will occur, and the lower the reading, the better are the chances for inclement weather. This situation occurs because surface high-pressure areas are associated with sinking air and normally fair weather, whereas surface low-pressure areas are associated with rising air and usually cloudy, wet weather. A steady rise in atmospheric pressure (a



● FIGURE 8.7 The aneroid barometer.



● FIGURE 8.8 A recording barograph.

rising barometer) usually indicates clearing weather or fair weather, whereas a steady drop in atmospheric pressure (a falling barometer) often signals the approach of a storm with inclement weather.

The *altimeter* and *barograph* are two types of aneroid barometers. Altimeters are aneroid barometers that measure pressure, but are calibrated to indicate altitude. Barographs are recording aneroid barometers. Basically, the barograph consists of a pen attached to an indicating arm that marks a continuous record of pressure on chart paper. The chart paper is attached to a drum rotated slowly by an internal mechanical clock (see ● Fig. 8.8).

PRESSURE READINGS The seemingly simple task of reading the height of the mercury column to obtain the air pressure is actually not all that simple. Being a fluid, mercury is sensitive to changes in temperature; it will expand when heated and contract when cooled. Consequently, to obtain accurate pressure readings without the influence of temperature, all mercury barometers are corrected as if they were read at the same temperature. Because the earth is not a perfect sphere, the force of gravity is not a constant. Since small gravity differences influence the height of the mercury column, they must be considered when reading the barometer. Finally, each barometer has its own "built-in" error, called *instrument error,* which is caused, in part, by the surface tension of the mercury against the glass tube. After being corrected for temperature, gravity, and instrument error, the barometer reading at a particular location and elevation is termed **station pressure.**

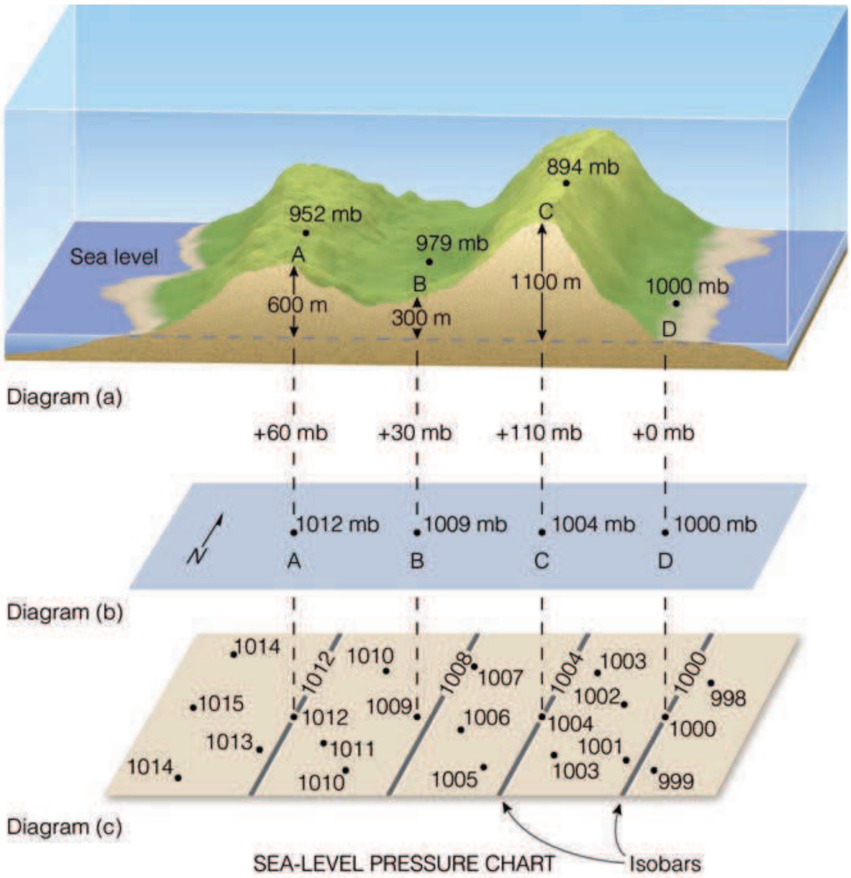
● Figure 8.9a gives the station pressure measured at four locations only a few hundred kilometers apart. The different station pressures of the four cities are due primarily to the cities being at different altitudes. This fact becomes even clearer when we realize that atmospheric pressure changes much more quickly when we move upward than it does when we move sideways. As an example, the vertical change in air pressure from the base to the top of the Empire State Building—a distance of a little more than ½ km—is typically much greater than the horizontal difference in air pressure from New York City to Miami, Florida—a distance of over 1600 km. Therefore, we can see that a small vertical difference between two observation sites can yield a large difference in station pressure. Thus, to properly monitor horizontal changes in pressure, barometer readings must be corrected for altitude.

Altitude corrections are made so that a barometer reading taken at one elevation can be compared with a barometer reading taken at another. Station pressure observations are normally adjusted to a level of *mean sea level*—the level representing the average surface of the ocean. The adjusted reading is called **sea-level pressure.** The size of the correction depends primarily on how high the station is above sea level.

Near the earth's surface, atmospheric pressure decreases on the average by about 10 mb for every 100 m increase in elevation (about 1 in. of mercury for each 1000-ft rise).* Notice in Fig. 8.9a that city A has a station pressure of 952 mb. Notice also that city A is 600 m above sea level. Adding 10 mb per 100 m to its station pressure yields a sea-level pressure of 1012 mb (Fig. 8.9b). After all the station pressures are adjusted to sea level (Fig. 8.9c), we are able to see the horizontal variations in sea-level pressure—something we were not able to see from the station pressures alone in Fig. 8.9a.

When more pressure data are added (see Fig. 8.9c), the chart can be analyzed and the pressure pattern visualized.

*This decrease in atmospheric pressure with height (10 mb/100 m) occurs when the air temperature decreases at the standard lapse rate of 6.5°C/1000 m. Because atmospheric pressure decreases more rapidly with height in cold (more-dense) air than it does in warm (less-dense) air, the vertical rate of pressure change is typically greater than 10 mb per 100 m in cold air and less than that in warm air.



● **FIGURE 8.9** The top diagram (a) shows four cities (A, B, C, and D) at varying elevations above sea level, all with different station pressures. The middle diagram (b) represents sea-level pressures of the four cities plotted on a sea-level chart. The bottom diagram (c) shows sea-level pressure readings of the four cities plus other sea-level pressure readings, with isobars drawn on the chart (gray lines) at intervals of 4 millibars.

Isobars (lines connecting points of equal pressure) are drawn at intervals of 4 mb,* with 1000 mb being the base value. Note that the isobars do not pass through each point, but, rather, between many of them, with the exact values being interpolated from the data given on the chart. For example, follow the 1008-mb line from the top of the chart southward and observe that there is no plotted pressure of 1008 mb. The 1008-mb isobar, however, comes closer to the station with a sea-level pressure of 1007 mb than it does to the station with a pressure of 1010 mb. With its isobars, the bottom chart (Fig. 8.9c) is now called a *sea-level pressure chart* or simply a **surface map.** When weather data are plotted on the map it becomes a *surface weather map.*

Surface and Upper-Level Charts

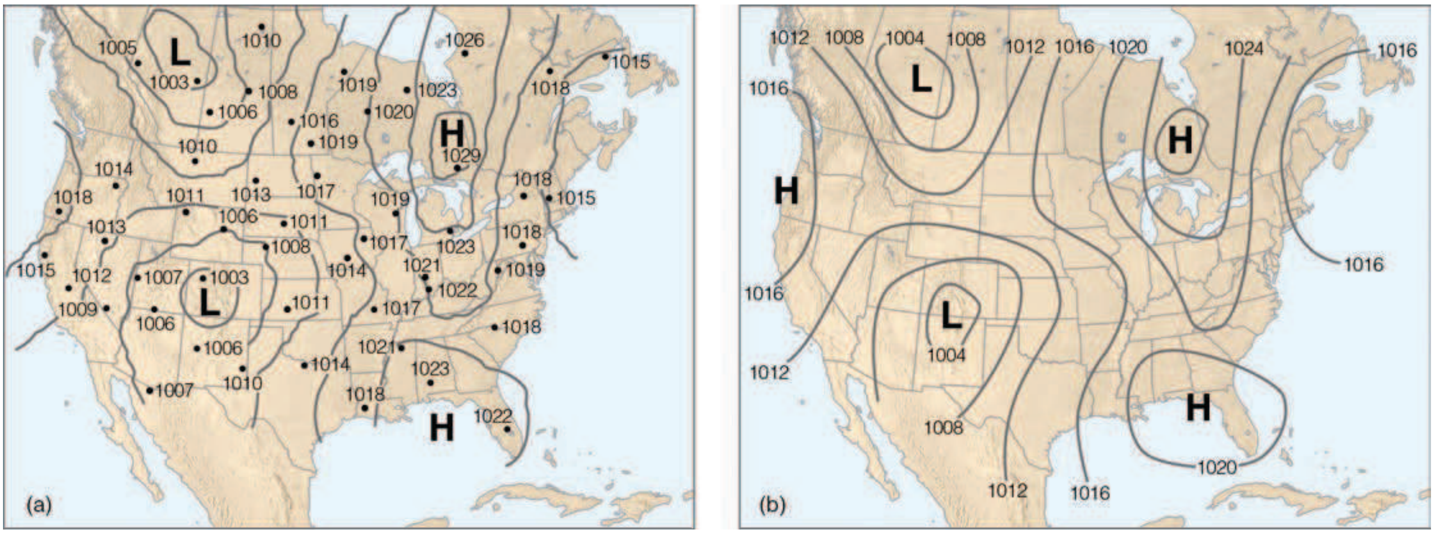
The isobars on the surface map in ● Fig. 8.10a are drawn precisely, with each individual observation taken into account. Notice that many of the lines are irregular, especially in mountainous regions over the Rockies. The reason for the wiggle is due, in part, to small-scale local variations in pressure and to errors introduced by correcting observations that were taken at high-altitude stations. An extreme case of this type of error occurs at Leadville, Colorado (elevation 3096 m), the highest city in the United States. Here, the station pressure is typically near 700 mb. This means that nearly 300 mb must be added to obtain a sea-level pressure reading! A mere 1 percent error in estimating the exact correction would result in a 3-mb error in sea-level pressure. For this reason, isobars are smoothed through readings from high-altitude stations and from stations that might have small observational errors. Figure 8.10b shows how the isobars appear on the surface map after they are smoothed.

The sea-level pressure chart described so far is called a *constant height chart* because it represents the atmospheric pressure at a constant level—in this case, sea level. The same type of chart could be drawn to show the horizontal variations in pressure at any level in the atmosphere; for example, at 3000 m (see ● Fig. 8.11).

Another type of chart commonly used in studying the weather is the *constant pressure chart,* or **isobaric chart.** Instead of showing pressure variations at a constant altitude, these charts are constructed to show height variations along an equal pressure *(isobaric)* surface. Constant pressure charts are convenient to use because the height variables they show are easier to deal with in meteorological equations than the variables of pressure. Since isobaric charts are in common use, let's examine them in detail.

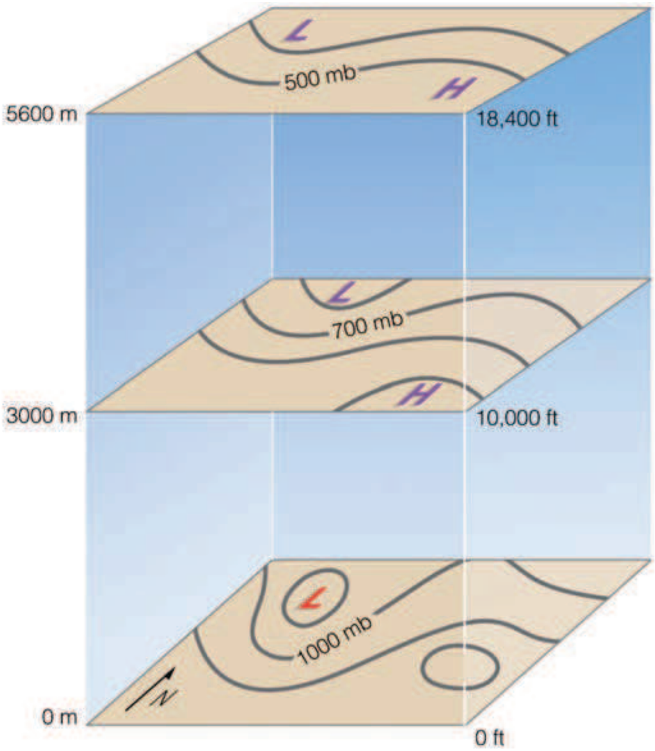
Imagine that the dots inside the air column in ● Fig. 8.12 represent tightly packed air molecules from the surface up to

*An interval of 2 mb would put the lines too close together, and an 8-mb interval would spread them too far apart.

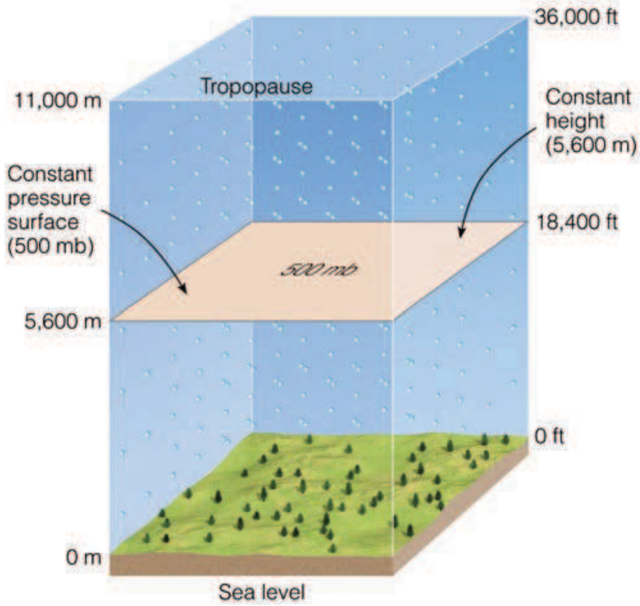


● FIGURE 8.10 (a) Sea-level isobars drawn so that each observation is taken into account. Not all observations are plotted. (b) Sea-level isobars after smoothing.

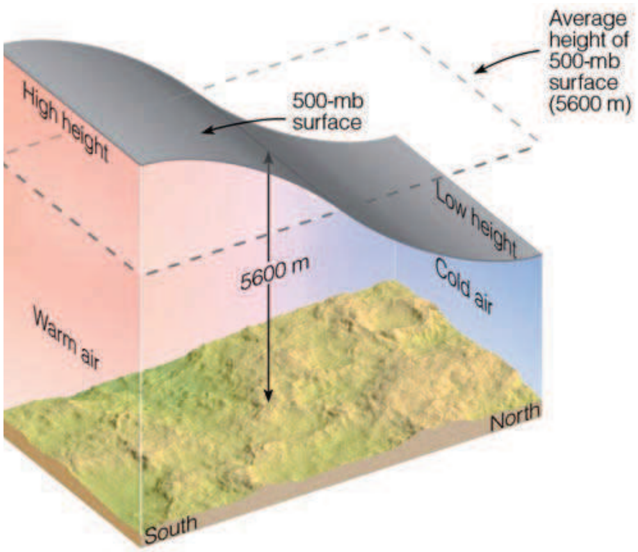
the tropopause. Assume that the air density is constant throughout the entire air layer and that all of the air molecules are squeezed into this layer. If we climb halfway up the air column and stop, then draw a sheetlike surface representing this level, we will have made a constant height surface. This altitude (5600 m) is where we would, under standard conditions, measure a pressure of 500 mb. Observe that everywhere along this surface (shaded tan in the diagram) there are an equal number of molecules above it. This condition means that the level of constant height also represents a level of constant pressure. At every point on this *isobaric surface,* the height is 5600 m above sea level and the pressure is 500 mb. Within the air column, we could cut any number of horizontal slices, each one at a different altitude, and each slice would represent both an isobaric and constant height surface. A map of any one of these surfaces would be blank, since there are no horizontal variations in either pressure or altitude.



● FIGURE 8.11 Each map shows isobars on a constant height chart. The isobars represent variations in horizontal pressure at that altitude. An average isobar at sea level would be about 1000 mb; at 3000 m, about 700 mb; and at 5600 m, about 500 mb.



● FIGURE 8.12 When there are no horizontal variations in pressure, constant pressure surfaces are parallel to constant height surfaces. In the diagram, a measured pressure of 500 mb is 5600 m above sea level everywhere. (Dots in the diagram represent air molecules.)



● FIGURE 8.13 The area shaded gray in the above diagram represents a surface of constant pressure, or isobaric surface. Because of the changes in air density, the isobaric surface rises in warm, less-dense air and lowers in cold, more-dense air. Where the horizontal temperature changes most quickly, the isobaric surface changes elevation most rapidly.

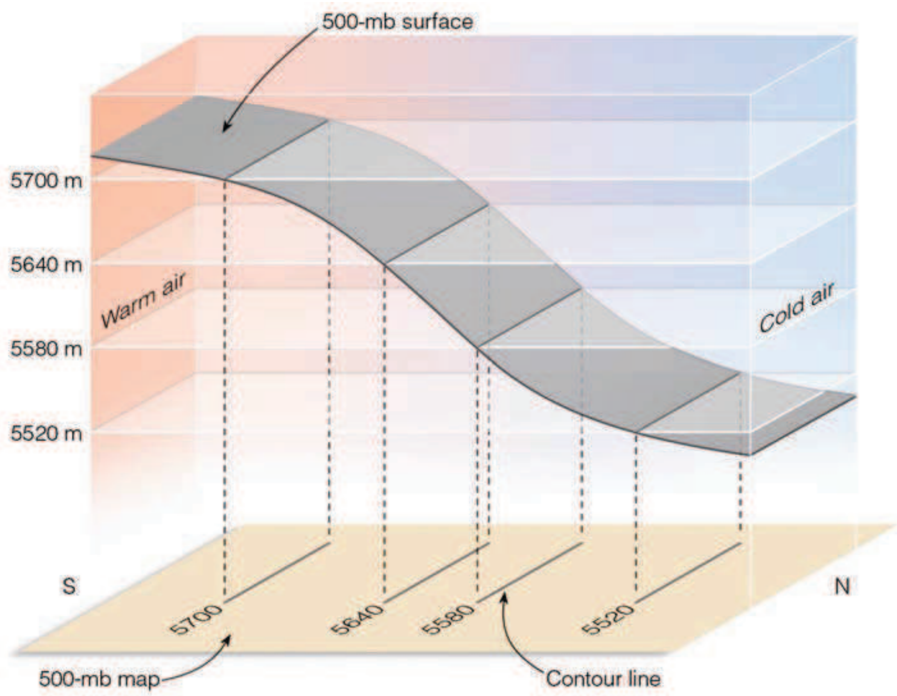
If the air temperature should change in any portion of the column, the air density and pressure would change along with it. Notice in ● Fig. 8.13 that we have colder air to the north and warmer air to the south. To simplify this situation, we will assume that the atmospheric pressure at the earth's surface remains constant. Hence, the total number of molecules in the column above each region must remain constant.

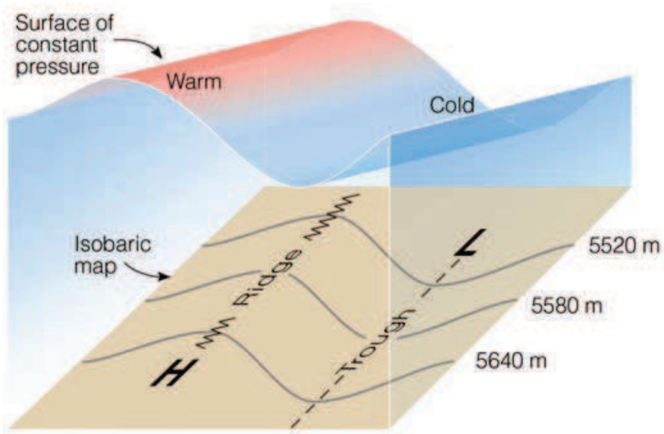
In Fig. 8.13, the area shaded gray at the top of the column represents a constant pressure (isobaric) surface, where the atmospheric pressure at all points along this surface is 500 mb. Notice that in the warmer, less-dense air the 500-mb pressure surface is found at a higher (than average) level, while in the colder, more-dense air, it is observed at a much lower (than average) level. From these observations, we can see that when the air aloft is warm, constant pressure surfaces are typically found at higher elevations than normal, and when the air aloft is cold, constant pressure surfaces are typically found at lower elevations than normal.

Look again at Fig. 8.13 and observe that in the warm air at an altitude of 5600 m, the atmospheric pressure must be greater than 500 mb, whereas in the cold air, at the same altitude (5600 m), the atmospheric pressure must be less than 500 mb. Therefore, we can conclude that *high heights on an isobaric chart correspond to higher-than-normal pressures at any given altitude, and low heights on an isobaric chart correspond to lower-than-normal pressures.*

The variations in height of the isobaric surface in Fig. 8.13 are shown in ● Fig. 8.14. Note that where the constant altitude lines intersect the 500-mb pressure surface, **contour lines** (lines connecting points of equal elevation) are drawn on the 500-mb map. Each contour line, of course, tells us the altitude above sea level at which we can obtain a pressure reading of 500 mb. In the warmer air to the south, the elevations are high, while in the cold air to the north, the elevations are low. The contour lines are crowded together in the middle of the chart, where the pressure surface dips rapidly due to the changing air temperatures. Where there is little horizontal temperature change, there are also few contour lines. Although contour lines are height lines, keep in mind that they illustrate pressure as do isobars in that contour lines of low height represent a region of lower pressure and contour lines of high height represent a region of higher pressure.

● FIGURE 8.14 Changes in elevation of an isobaric surface (500 mb) show up as contour lines on an isobaric (500 mb) map. Where the surface dips most rapidly, the lines are closer together.





● FIGURE 8.15 The wavelike patterns of an isobaric surface reflect the changes in air temperature. An elongated region of warm air aloft shows up on an isobaric map as higher heights and a ridge; the colder air shows as lower heights and a trough.

Since cold air aloft is normally associated with low heights or low pressures, and warm air aloft with high heights or high pressures, on upper-air charts representing the Northern Hemisphere, contour lines and isobars usually decrease in value from south to north because the air is typically warmer to the south and colder to the north. The lines, however, are not straight; they bend and turn, indicating **ridges** *(elongated highs)* where the air is warm and indicating depressions, or **troughs** *(elongated lows)*, where the air is cold. In ● Fig. 8.15, we can see how the wavy contours on the map relate to the changes in altitude of the isobaric surface.

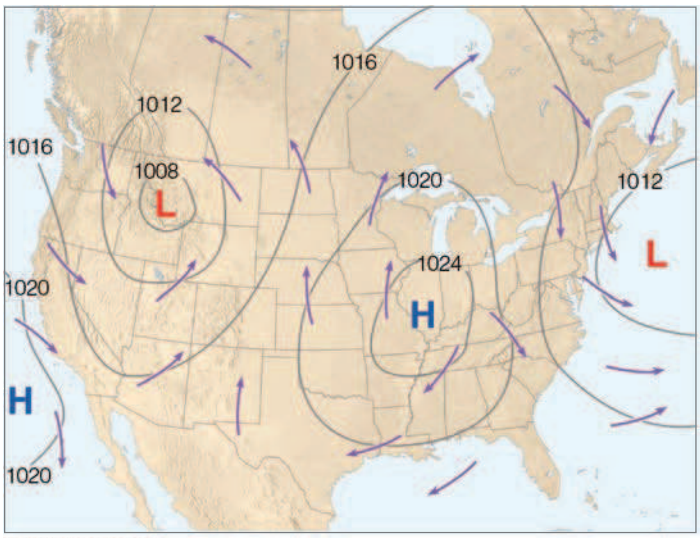
Although we have examined only the 500-mb chart, other isobaric charts are commonly used. ▼ Table 8.1 lists these charts and their approximate heights above sea level.

▼ TABLE 8.1 Common Isobaric Charts and Their Approximate Elevation above Sea Level

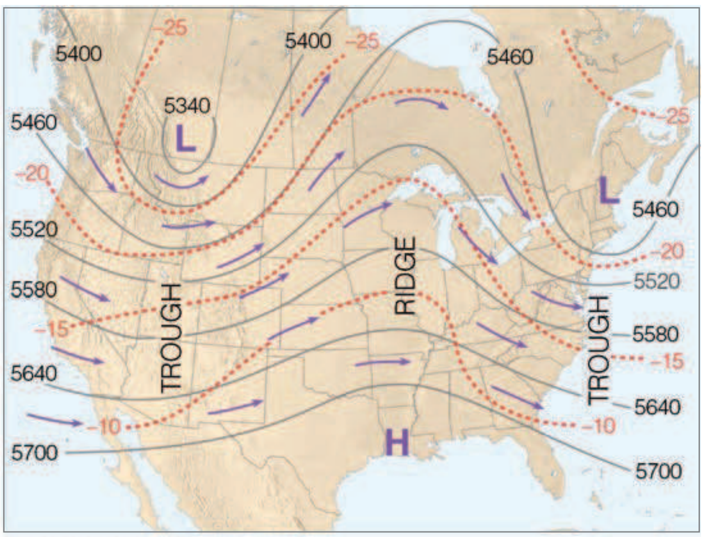
ISOBARIC SURFACE (MB) CHARTS	APPROXIMATE ELEVATION (M)	(FT)
1000	120	400
850	1,460	4,800
700	3,000	9,800
500	5,600	18,400
300	9,180	30,100
200	11,800	38,700
100	16,200	53,200

Upper-level charts are a valuable tool. As we will see, they show wind-flow patterns that are extremely important in forecasting the weather. They can also be used to determine the movement of weather systems and to predict the behavior of surface pressure areas. To the pilot of a small aircraft, a constant pressure chart can help determine whether the plane is flying at an altitude either higher or lower than its altimeter indicates. (For more information on this topic, read the Focus section "Flying on a Constant Pressure Surface—High to Low, Look Out Below," p. 204.)

● Figure 8.16a is a simplified surface map that shows areas of high and low pressure and arrows that indicate *wind direction*—the direction from which the wind is blowing. The large blue H's on the map indicate the centers of high pressure, which are also called **anticyclones.** The large L's represent centers of low pressure, also known as *depressions* or **mid-latitude cyclonic storms** because they form in the



(a) Surface map



(b) Upper-air map (500 mb)

● FIGURE 8.16 (a) Surface map showing areas of high and low pressure. The solid lines are isobars drawn at 4-mb intervals. The arrows represent wind direction. Notice that the wind blows *across* the isobars. (b) The upper-level (500-mb) map for the same day as the surface map. Solid lines on the map are contour lines in meters above sea level. Dashed red lines are isotherms in °C. Arrows show wind direction. Notice that, on this upper-air map, the wind blows *parallel* to the contour lines.

FOCUS ON AN OBSERVATION

Flying on a Constant Pressure Surface—High to Low, Look Out Below

Aircraft that use pressure altimeters typically fly along a constant pressure surface rather than a constant altitude surface. They do this because the *altimeter,* as we saw earlier, is simply an aneroid barometer calibrated to convert atmospheric pressure to an approximate elevation. The altimeter elevation indicated by an altimeter assumes a standard atmosphere where the air temperature decreases at the rate of 6.5°C/1000 m (3.6°F/1000 ft). Since the air temperature seldom, if ever, decreases at exactly this rate, altimeters generally indicate an altitude different from their true elevation.

Figure 2 shows a standard column of air bounded on each side by air with a different temperature and density. On the left side, the air is warm; on the right, it is cold. The orange line represents a constant pressure surface of 700 mb as seen from the side. In the standard air, the 700-mb surface is located at 10,000 ft above sea level.

In the warm air, the 700-mb surface rises; in the cold air, it descends. An aircraft flying along the 700-mb surface would be at an altitude less than 10,000 ft in the cold air, equal to 10,000 ft in the standard air, and greater than 10,000 ft in the warmer air. With no corrections for temperature, the altimeter would indicate the same altitude at all three positions because the air pressure does not change. We can see that, if no temperature corrections are made, an aircraft flying into warm air will increase in altitude and fly higher than its altimeter indicates. Put another way: The altimeter inside the plane will read an altitude lower than the plane's true elevation.

Flying from standard air into cold air represents a potentially dangerous situation. As an aircraft flies into cold air, it flies along a lowering pressure surface. If no correction for temperature is made, the altimeter shows no change in elevation even though the aircraft is losing altitude; hence, the plane will be flying lower than the altimeter indicates. This problem can be serious, especially for planes flying above mountainous terrain with poor visibility and where high winds and turbulence can reduce the air pressure drastically. To ensure adequate clearance under these conditions, pilots fly their aircraft higher than they normally would, consider air temperature, and compute a more realistic altitude by resetting their altimeters to reflect these conditions.

Even without sharp temperature changes, pressure surfaces may dip suddenly (see Fig. 3). An aircraft flying into an area of decreasing pressure will lose altitude unless corrections are made. For example, suppose a pilot has set the altimeter for sea-level pressure above station A. At this location, the plane is flying along an isobaric surface at a true altitude of 500 ft. As the plane flies toward station B, the pressure surface (and the plane) dips but the altimeter continues to read 500 ft, which is too high. To correct for such changes in pressure, a pilot can obtain a current altimeter setting from ground control. With this additional information, the altimeter reading will more closely match the aircraft's actual altitude.

Because of the inaccuracies inherent in the pressure altimeter, many high performance and commercial aircraft are equipped with a radio altimeter. This device is like a small radar unit that measures the altitude of the aircraft by sending out radio waves, which bounce off the terrain. The time it takes these waves to reach the surface and return is a measure of the aircraft's altitude. If used in conjunction with a pressure altimeter, a pilot can determine the variations in a constant pressure surface simply by flying along that surface and observing how the true elevation measured by the radio altimeter changes.

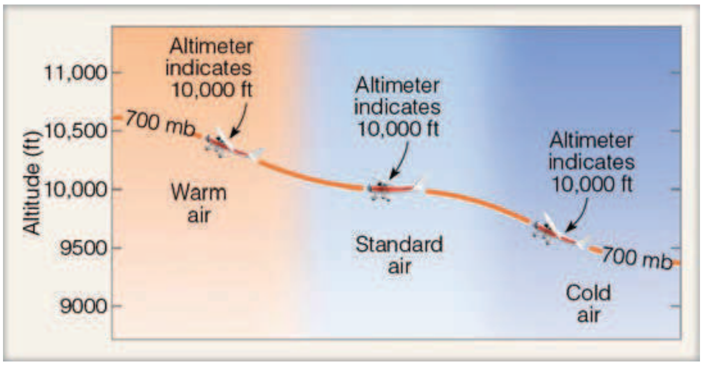


FIGURE 2 An aircraft flying along a surface of constant pressure (orange line) may change altitude as the air temperature changes. Without being corrected for the temperature change, a pressure altimeter will continue to read the same elevation.

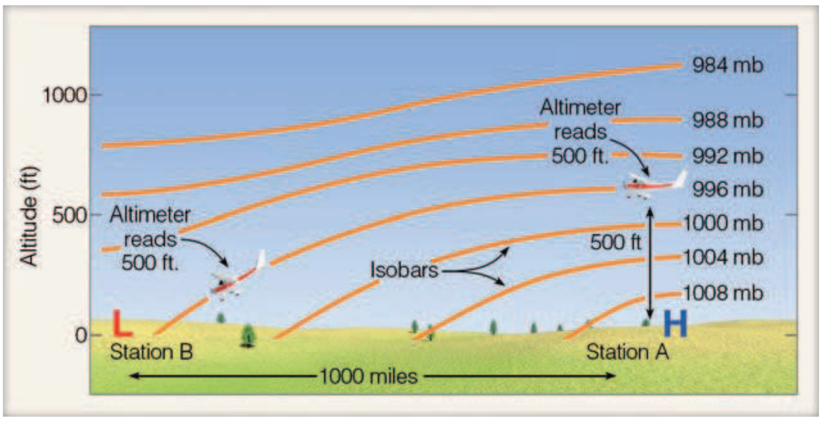


FIGURE 3 In the absence of horizontal temperature changes, pressure surfaces can dip toward the earth's surface. An aircraft flying along the pressure surface will either lose or gain altitude, depending on the direction of flight.

middle latitudes, outside of the tropics. The solid dark lines are isobars with units in millibars. Notice that the surface winds tend to blow across the isobars toward regions of lower pressure. In fact, as we briefly observed in Chapter 1, in the Northern Hemisphere the winds blow counterclockwise and inward toward the center of the lows and clockwise and outward from the center of the highs.

Figure 8.16b shows an upper-air chart (a 500-mb isobaric map) for the same day as the surface map in Fig. 8.16a. The solid gray lines on the map are contour lines given in meters above sea level. The difference in elevation between each contour line (called the *contour interval*) is 60 meters. Superimposed on this map are dashed red lines, which represent lines of equal temperature (isotherms). Observe how the contour lines tend to parallel the isotherms. As we would expect, the contour lines tend to decrease in value from south to north.

The arrows on the 500-mb map show the wind direction. Notice that, unlike the surface winds that cross the isobars in Fig. 8.16a, the winds on the 500-mb chart tend to flow *parallel* to the contour lines in a wavy west-to-east direction. Why does the wind tend to cross the isobars on a surface map, yet blow parallel to the contour lines (or isobars) on an upper-air chart? To answer this question we will now examine the forces that affect winds.

Newton's Laws of Motion

Our understanding of why the wind blows stretches back through several centuries, with many scientists contributing to our knowledge. When we think of the movement of air, however, one great scholar stands out—Isaac Newton (1642–1727), who formulated several fundamental laws of motion.

Newton's first law of motion states that *an object at rest will remain at rest and an object in motion will remain in motion (and travel at a constant velocity along a straight line) as long as no force is exerted on the object.* For example, a baseball in a pitcher's hand will remain there until a force (a push) acts upon the ball. Once the ball is pushed (thrown), it would continue to move in that direction forever if it were not for the force of air friction (which slows it down), the force of gravity (which pulls it toward the ground), and the catcher's mitt (which exerts an equal but opposite force to bring it to a halt). Similarly, to start air moving, to speed it up, to slow it down, or even to change its direction requires the action of an external force. This brings us to Newton's second law.

Newton's second law states that *the force exerted on an object equals its mass times the acceleration produced.* In symbolic form, this law is written as

$$F = ma.$$

From this relationship we can see that, when the mass of an object is constant, the force acting on the object is directly related to the acceleration that is produced. A force in its simplest form is a push or a pull. *Acceleration is the speeding up, the slowing down, or the changing of direction of an object.* (More precisely, acceleration is the change in velocity* over a period of time.)

Because more than one force may act upon an object, Newton's second law always refers to the *net,* or total, force that results. An object will always accelerate in the direction of the total force acting on it. Therefore, to determine in which direction the wind will blow, we must identify and examine all of the forces that affect the horizontal movement of air. These forces include:

1. pressure gradient force
2. Coriolis force
3. centripetal force
4. friction

We will first study the forces that influence the flow of air aloft. Then we will see which forces modify winds near the ground.

Forces That Influence the Winds

We already know that horizontal differences in atmospheric pressure cause air to move and, hence, the wind to blow. Since air is an invisible gas, it may be easier to see how pressure differences cause motion if we examine a visible fluid, such as water.

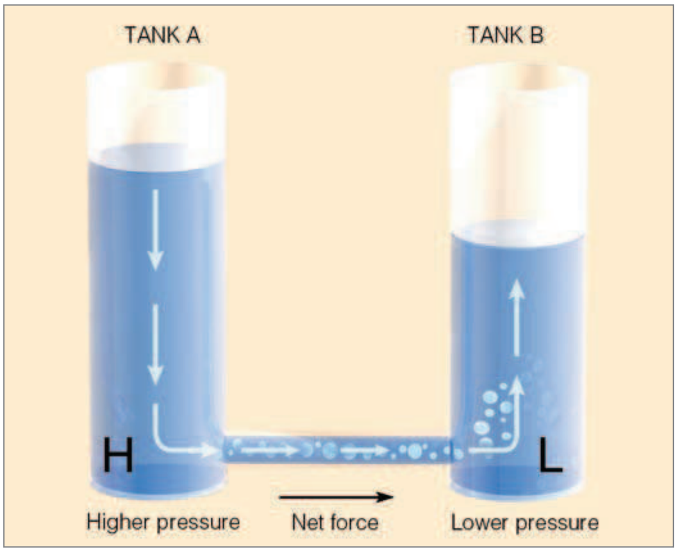
In ● Fig. 8.17, the two large tanks are connected by a pipe. Tank A is two-thirds full and tank B is only one-half full. Since the water pressure at the bottom of each tank is proportional to the weight of water above, the pressure at the bottom of tank A is greater than the pressure at the bottom of tank B. Moreover, since fluid pressure is exerted equally in all directions, there is a greater pressure in the pipe directed from tank A toward tank B than from B toward A.

Since pressure is force per unit area, there must also be a net force directed from tank A toward tank B. This force causes the water to flow from left to right, from higher pressure toward lower pressure. The greater the pressure difference, the stronger the force, and the faster the water moves. In a similar way, horizontal differences in atmospheric pressure cause air to move.

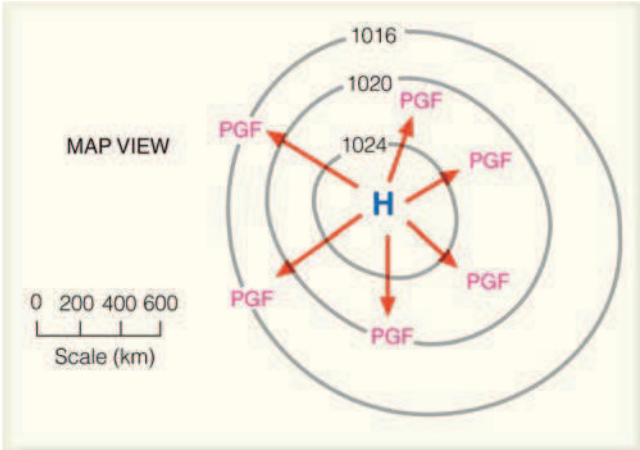
PRESSURE GRADIENT FORCE ● Figure 8.18 shows a region of higher pressure on the map's left side, lower pressure on the right. The isobars show how the horizontal pressure is changing. If we compute the amount of pressure change that occurs over a given distance, we have the **pressure gradient;** thus

$$\text{Pressure gradient} = \frac{\text{difference in pressure}}{\text{distance}}.$$

*Velocity specifies both the speed of an object and its direction of motion.



● FIGURE 8.17 The higher water level creates higher fluid pressure at the bottom of tank A and a net force directed toward the lower fluid pressure at the bottom of tank B. This net force causes water to move from higher pressure toward lower pressure.



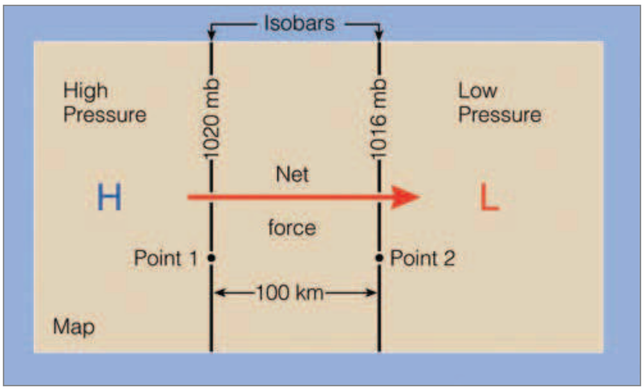
● FIGURE 8.19 The closer the spacing of the isobars, the greater the pressure gradient. The greater the pressure gradient, the stronger the pressure gradient force *(PGF)*. The stronger the *PGF*, the greater the wind speed. The red arrows represent the relative magnitude of the force, which is always directed from higher toward lower pressure.

If we let the symbol delta (Δ) mean "a change in," we can simplify the expression and write the pressure gradient as

$$PG = \frac{\Delta p}{d},$$

where Δp is the pressure difference between two places some horizontal distance *(d)* apart. In Fig. 8.18 the pressure gradient between points 1 and 2 is 4 mb per 100 km.

Suppose the pressure in Fig. 8.18 were to change and the isobars become closer together. This condition would produce a rapid change in pressure over a relatively short distance, or what is called a *steep* (or *strong*) *pressure gradient.* However, if the pressure were to change such that the isobars spread farther apart, then the difference in pressure would be small over a relatively large distance. This condition is called a *gentle* (or *weak*) *pressure gradient.*



● FIGURE 8.18 The pressure gradient between point 1 and point 2 is 4 mb per 100 km. The net force directed from higher toward lower pressure is the *pressure gradient force.*

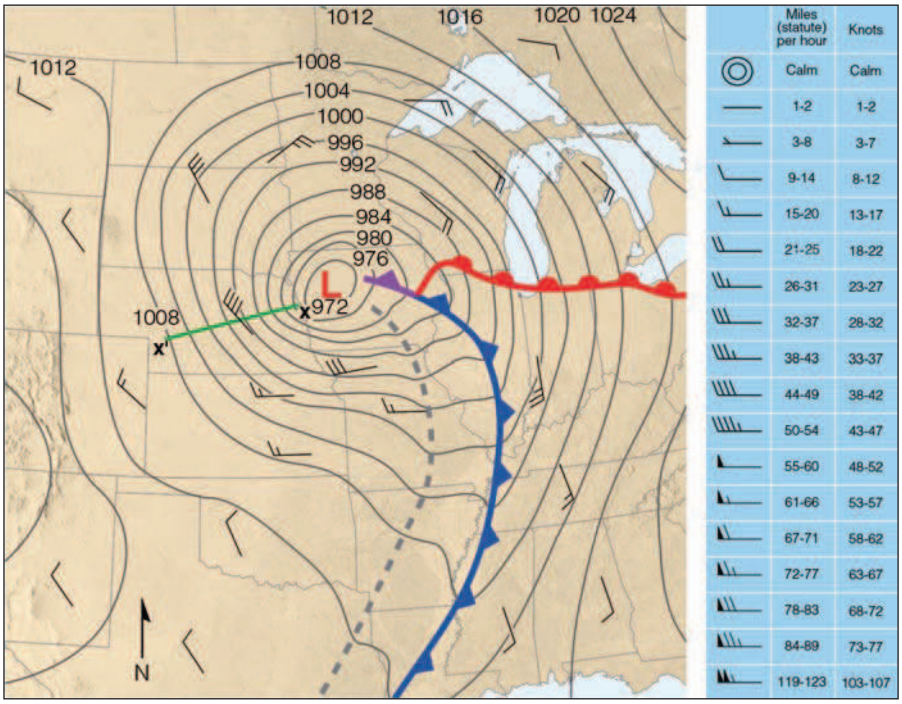
Notice in Fig. 8.18 that when differences in horizontal air pressure exist there is a net force acting on the air. This force, called the **pressure gradient force** *(PGF), is directed from higher toward lower pressure at right angles to the isobars.* The magnitude of the force is directly related to the pressure gradient. Steep pressure gradients correspond to strong pressure gradient forces and vice versa. ● Figure 8.19 shows the relationship between pressure gradient and pressure gradient force.

The *pressure gradient force is the force that causes the wind to blow.* Because of this effect, closely spaced isobars on a weather map indicate steep pressure gradients, strong forces, and high winds. On the other hand, widely spaced isobars indicate gentle pressure gradients, weak forces, and light winds. An example of a steep pressure gradient and strong winds is given in ● Fig. 8.20. Notice that the tightly packed isobars along the green line are producing a steep pressure gradient of 32 mb per 500 km and strong surface winds of 40 knots.

If the pressure gradient force were the only force acting upon air, we would always find winds blowing directly from higher toward lower pressure. However, the moment air starts to move, it is deflected in its path by the *Coriolis force.*

CORIOLIS FORCE The Coriolis force describes an apparent force that is due to the rotation of the earth. To understand how it works, consider two people playing catch as they sit opposite one another on the rim of a merry-go-round (see ● Fig. 8.21, platform A). If the merry-go-round is not moving, each time the ball is thrown, it moves in a straight line to the other person.

Suppose the merry-go-round starts turning counterclockwise—the same direction the earth spins as viewed from above the North Pole. If we watch the game of catch from

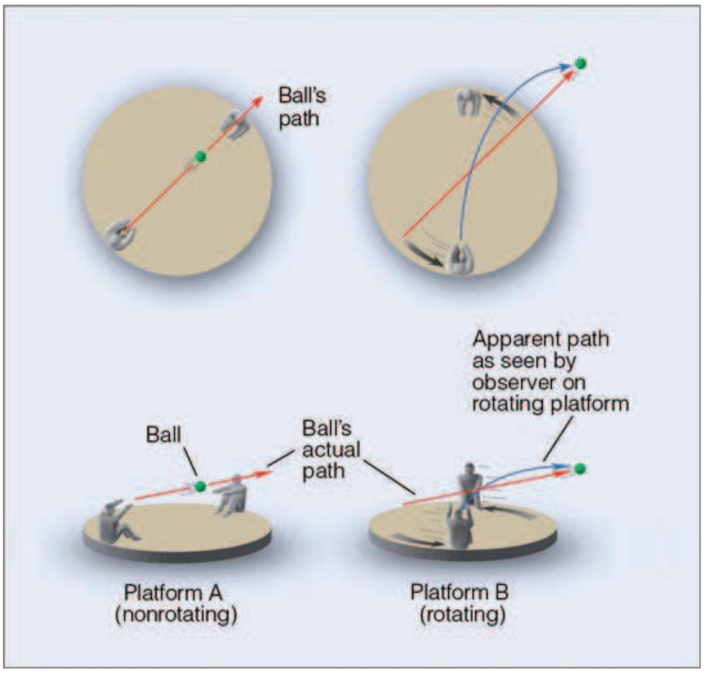


● FIGURE 8.20 Surface weather map for 6 A.M. (CST), Tuesday, November 10, 1998. Dark gray lines are isobars with units in millibars. The interval between isobars is 4 mb. A deep low with a central pressure of 972 mb (28.70 in.) is moving over northwestern Iowa. The distance along the green line X-X′ is 500 km. The difference in pressure between X and X′ is 32 mb, producing a pressure gradient of 32 mb/500 km. The tightly packed isobars along the green line are associated with strong northwesterly winds of 40 knots, with gusts even higher. Wind directions are given by lines that parallel the wind. Wind speeds are indicated by barbs and flags. (A wind indicated by the symbol would be a wind from the northwest at 10 knots. See blue insert.) The solid blue line is a cold front, the solid red line a warm front, and the solid purple line an occluded front. The dashed gray line is a trough.

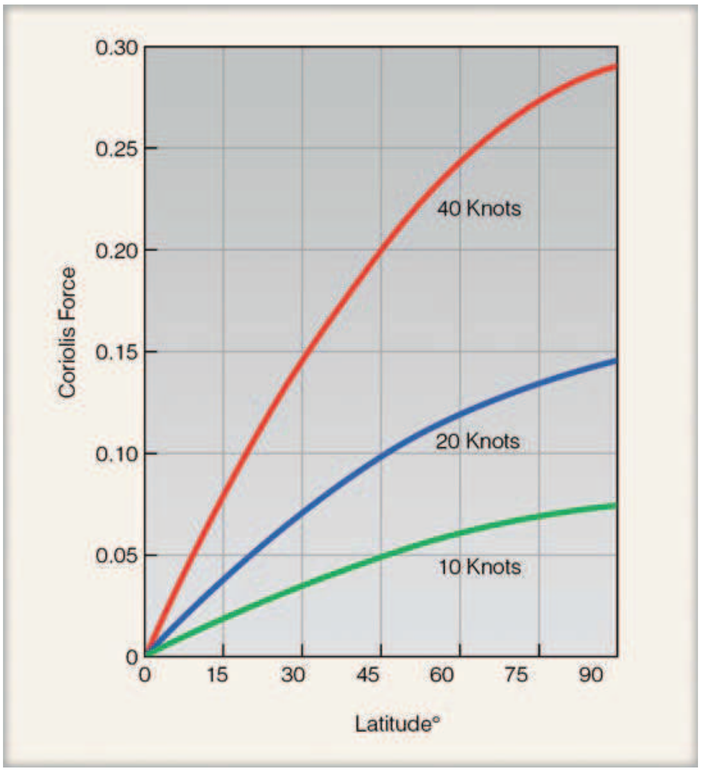
above, we see that the ball moves in a straight-line path just as before. However, to the people playing catch on the merry-go-round, the ball seems to veer to its right each time it is thrown, always landing to the right of the point intended by the thrower (see Fig. 8.21, platform B). This perception is due to the fact that, while the ball moves in a straight-line path, the merry-go-round rotates beneath it; by the time the ball reaches the opposite side, the catcher has moved. To anyone on the merry-go-round, it seems as if there is some force causing the ball to deflect to the right. This apparent force is called the **Coriolis force** after Gaspard Coriolis, a nineteenth-century French scientist who worked it out mathematically. (Because it is an *apparent force* due to the rotation of the earth, it is also called the *Coriolis effect.*) This effect occurs on the rotating earth, too. All free-moving objects, such as ocean currents, aircraft, artillery projectiles, and air molecules seem to deflect from a straight-line path because the earth rotates under them.

The Coriolis force *causes the wind to deflect to the right of its intended path in the Northern Hemisphere and to the left of its intended path in the Southern Hemisphere.* To illustrate this, consider a satellite in polar circular orbit. If the earth were not rotating, the path of the satellite would be observed to move directly from north to south, parallel to the earth's meridian lines. However, the earth *does* rotate, carrying us and meridians eastward with it. Because of this rotation in the Northern Hemisphere, we see the satellite moving southwest instead of due south; it seems to veer off its path and move toward *its right.* In the Southern Hemisphere, the earth's direction of rotation is clockwise as viewed from above the South Pole. Consequently, a satellite moving northward from the South Pole would appear to move northwest and, hence, would veer to the *left* of its path.

The magnitude of the Coriolis force varies with the speed of the moving object and the latitude. ● Figure 8.22 shows this variation for various wind speeds at different latitudes. In each case, as the wind speed increases, the Coriolis force increases; hence, *the stronger the wind speed, the greater the de-*



● FIGURE 8.21 On nonrotating platform A, the thrown ball moves in a straight line. On platform B, which rotates counterclockwise, the ball continues to move in a straight line. However, platform B is rotating while the ball is in flight; thus, to anyone on platform B, the ball appears to deflect to the right of its intended path.



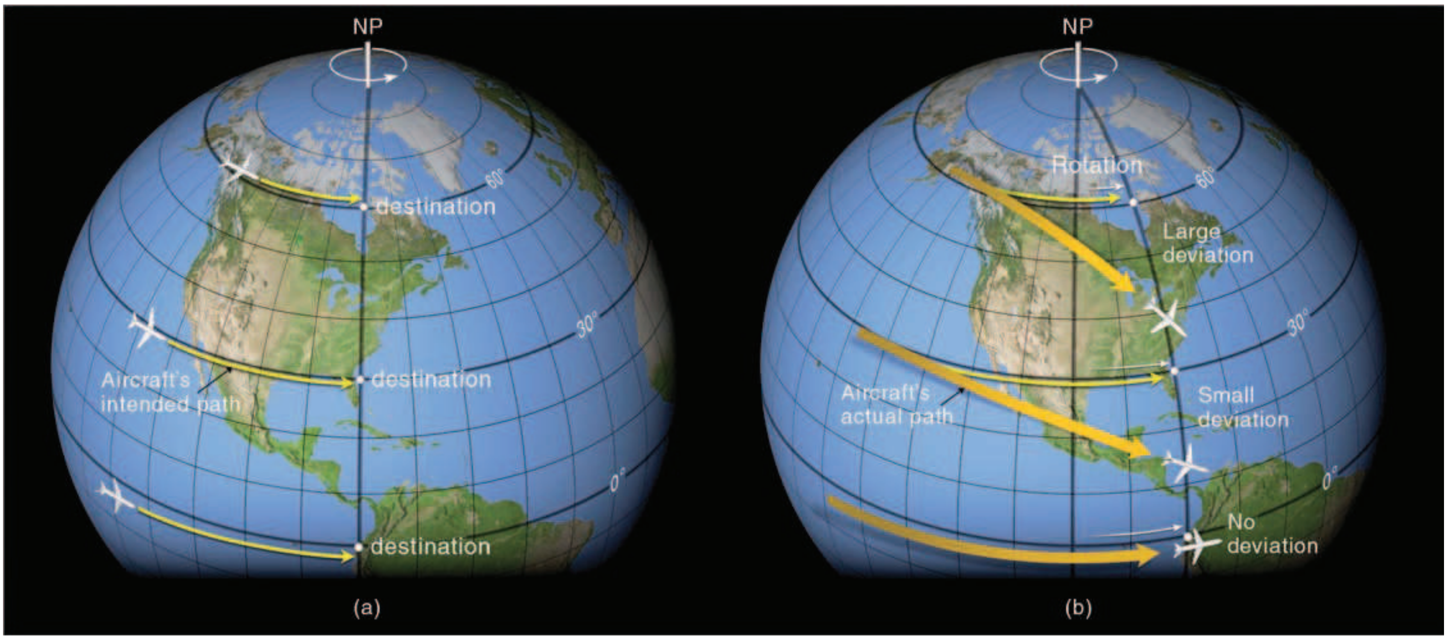
• FIGURE 8.22 The relative variation of the Coriolis force at different latitudes with different wind speeds.

WEATHER WATCH

The deep, low-pressure area illustrated in Fig. 8.20 was quite a storm. The intense low with its tightly packed isobars and strong pressure gradient produced extremely high winds that gusted over 90 knots in Wisconsin. The extreme winds caused blizzard conditions over the Dakotas, closed many interstate highways, shut down airports, and overturned trucks. The winds pushed a school bus off the road near Albert Lea, Minnesota, injuring two children, and blew the roofs off homes in Wisconsin. This notorious deep storm set an all-time record low pressure of 963 mb (28.43 in.) for Minnesota on November 10, 1998.

flection. Also, note that the Coriolis force increases for all wind speeds from a value of *zero at the equator to a maximum at the poles.* We can see this latitude effect better by examining • Fig. 8.23.

Imagine in Fig. 8.23 that there are three aircraft, each at a different latitude and each flying along a straight-line path, with no external forces acting on them. The destination of each aircraft is due east and is marked on the diagram (see Fig. 8.23a). Each plane travels in a straight path relative to an observer positioned at a fixed spot in space. The earth rotates beneath the moving planes, causing the destination points at latitudes 30° and 60° to change direction slightly (to the observer in space) (see Fig. 8.23b). To an observer standing on the earth, however, it is the plane that appears to deviate. The amount of deviation is greatest toward the pole and nonexistent at the equator. Therefore, the Coriolis force has a far



ACTIVE FIGURE 8.23 Except at the equator, a free-moving object heading either east or west (or any other direction) will appear from the earth to deviate from its path as the earth rotates beneath it. The deviation (Coriolis force) is greatest at the poles and decreases to zero at the equator. Visit the Meteorology Resource Center to view this and other active figures at academic.cengage.com/login greater effect on the plane at high latitudes (large deviation) than on the plane at low latitudes (small deviation). On the equator, it has no effect at all. The same, of course, is true of its effect on winds.

In summary, to an observer on the earth, objects moving in *any direction* (north, south, east, or west) are deflected to the *right* of their intended path in the Northern Hemisphere and to the *left* of their intended path in the Southern Hemisphere. The amount of deflection depends upon:

1. the rotation of the earth
2. the latitude
3. the object's speed*

In addition, the *Coriolis force acts at right angles to the wind, only influencing wind direction and never wind speed.*

The Coriolis "force" behaves as a real force, constantly tending to "pull" the wind to its right in the Northern Hemisphere and to its left in the Southern Hemisphere. Moreover, this effect is present in all motions relative to the earth's surface. However, in most of our everyday experiences, the Coriolis force is so small (compared to other forces involved in those experiences) that it is negligible and, contrary to popular belief, does not cause water to turn clockwise or counterclockwise when draining from a sink. The Coriolis force is also minimal on small-scale winds, such as those that blow inland along coasts in summer. Here, the Coriolis force might be strong because of high winds, but the force cannot produce much deflection over the relatively short distances. Only where winds blow over vast regions is the effect significant.

View this concept in action on the Meteorology Resource Center at academic.cengage.com/login

BRIEF REVIEW

In summary, we know that:

- Atmospheric (air) pressure is the pressure exerted by the mass of air above a region.
- A change in surface air pressure can be brought about by changing the mass (amount of air) above the surface.
- Heating and cooling columns of air can establish horizontal variations in atmospheric pressure aloft and at the surface.
- A difference in horizontal air pressure produces a horizontal pressure gradient force.
- The pressure gradient force is always directed from higher pressure toward lower pressure, and it is the pressure gradient force that causes the air to move and the wind to blow.
- Steep pressure gradients (tightly packed isobars on a weather map) indicate strong pressure gradient forces and high winds; gentle pressure gradients (widely spaced isobars) indicate weak pressure gradient forces and light winds.
- Once the wind starts to blow, the Coriolis force causes it to bend to the right of its intended path in the Northern Hemisphere and to the left of its intended path in the Southern Hemisphere.

WEATHER WATCH

As you drive your car along a highway (at the speed limit), the Coriolis force would "pull" your vehicle to the right about 1500 feet for every 100 miles you travel if it were not for the friction between your tires and the road surface.

With this information in mind, we will first examine how the pressure gradient force and the Coriolis force produce straight-line winds aloft. We will then see what influence the centripetal force has on winds that blow along a curved path.

STRAIGHT-LINE FLOW ALOFT—GEOSTROPHIC WINDS

Earlier in this chapter, we saw that the winds aloft on an upper-level chart blow more or less parallel to the isobars or contour lines. We can see why this phenomenon happens by carefully looking at ● Fig. 8.24, which shows a map in the Northern Hemisphere, above the earth's frictional influence,* with horizontal pressure variations at an altitude of about 1 km above the earth's surface. The evenly spaced isobars indicate a constant pressure gradient force *(PGF)* directed from south toward north as indicated by the red arrow at the left. Why, then, does the map show a wind blowing from the west? We can answer this question by placing a parcel of air at position 1 in the diagram and watching its behavior.

At position 1, the *PGF* acts immediately upon the air parcel, accelerating it northward toward lower pressure. However, the instant the air begins to move, the Coriolis force deflects the air toward its right, curving its path. As the parcel of air increases in speed (positions 2, 3, and 4), the magnitude of the Coriolis force increases (as shown by the longer arrows), bending the wind more and more to its right. Eventually, the wind speed increases to a point where the Coriolis force just balances the *PGF.* At this point (position 5), the wind no longer accelerates because the net force is zero. Here the wind flows in a straight path, parallel to the isobars at a constant speed.† This flow of air is called a **geostrophic**

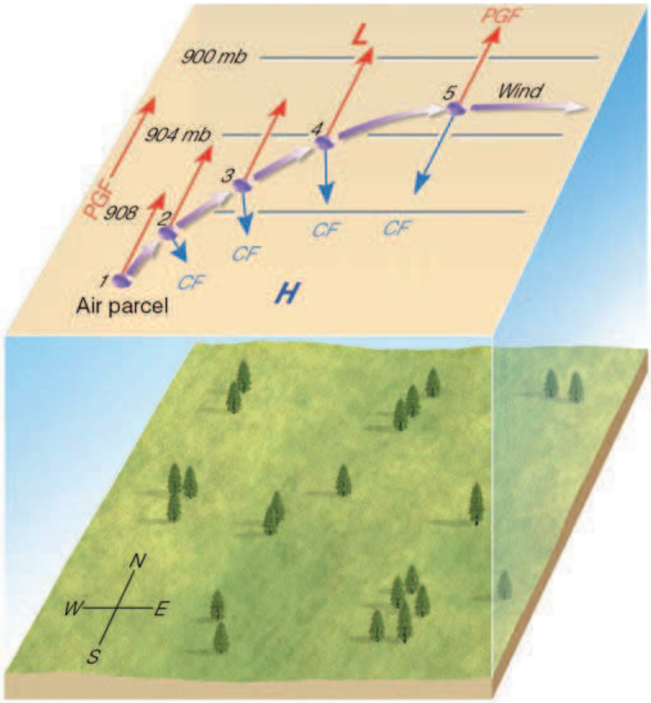
*These three factors are grouped together and shown in the expression

$$\text{Coriolis force} = 2m\,\Omega V \sin\phi,$$

where m is the object's mass, Ω is the earth's angular rate of spin (a constant), V is the speed of the object, and ϕ is the latitude.

*The friction layer (the layer where the wind is influenced by frictional interaction with objects on the earth's surface) usually extends from the surface up to about 1000 m (3300 ft) above the ground.

†At first, it may seem odd that the wind blows at a constant speed with no net force acting on it. But when we remember that the net force is necessary only to accelerate $(F = ma)$ the wind, it makes more sense. For example, it takes a considerable net force to push a car and get it rolling from rest. But once the car is moving, it only takes a force large enough to counterbalance friction to keep it going. There is no net force acting on the car, yet it rolls along at a constant speed.

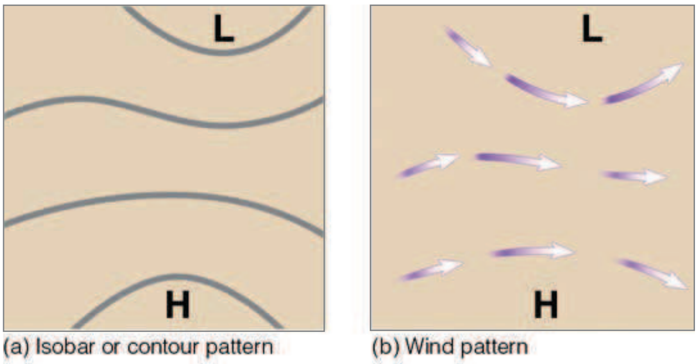


● FIGURE 8.24 Above the level of friction, air initially at rest will accelerate until it flows parallel to the isobars at a steady speed with the pressure gradient force *(PGF)* balanced by the Coriolis force *(CF)*. Wind blowing under these conditions is called geostrophic.

(*geo:* earth; *strophic:* turning) **wind.** Notice that the geostrophic wind blows in the Northern Hemisphere with lower pressure to its left and higher pressure to its right.

When the flow of air is purely geostrophic, the isobars (or contours) are straight and evenly spaced, and the wind speed is constant. In the atmosphere, isobars are rarely straight or evenly spaced, and the wind normally changes speed as it flows along. So, the geostrophic wind is usually only an approximation of the real wind. However, the approximation is generally close enough to help us more clearly understand the behavior of the winds aloft.

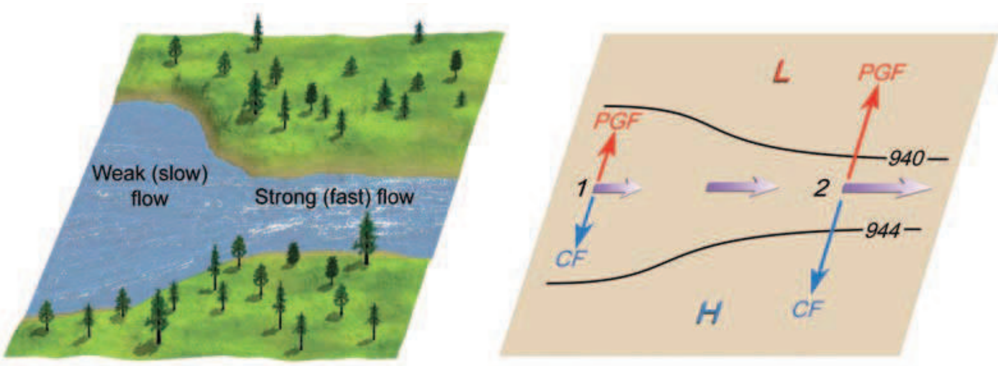
As we would expect from our previous discussion of winds, the speed of the geostrophic wind is directly related to the pressure gradient. In ● Fig. 8.25, we can see that a geostrophic wind flowing parallel to the isobars is similar to water in a stream flowing parallel to its banks. At position 1, the wind is blowing at a low speed; at position 2, the pressure gradient increases and the wind speed picks up. Notice also that at position 2, where the wind speed is greater, the Coriolis force is greater and balances the stronger pressure gradient force. (A more mathematical approach to the concept of geostrophic wind is given in the Focus section on p. 211.)



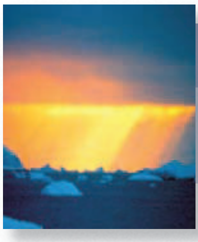
● FIGURE 8.26 By observing the orientation and spacing of the isobars (or contours) in diagram (a), the geostrophic wind direction and speed can be determined in diagram (b).

In ● Fig. 8.26, we can see that the geostrophic wind direction can be determined by studying the orientation of the isobars; its speed can be estimated from the spacing of the isobars. On an isobaric chart, the geostrophic wind direction and speed are related in a similar way to the contour lines. Therefore, if we know the isobar or contour patterns on an upper-level chart, we also know the direction and relative speed of the geostrophic wind, even for regions where no direct wind measurements have been made. Similarly, if we know the geostrophic wind direction and speed, we can estimate the orientation and spacing of the isobars, even if we don't have a current weather map. (It is also possible to estimate the wind flow and pressure patterns aloft by watching the movement of clouds. The Focus section on p. 212 illustrates this further.)

We know that the winds aloft do not always blow in a straight line; frequently, they curve and bend into meandering loops as they tend to follow the patterns of the isobars. In the Northern Hemisphere, winds blow counterclockwise around lows and clockwise around highs. The next section explains why.



● FIGURE 8.25 The isobars and contours on an upper-level chart are like the banks along a flowing stream. When they are widely spaced, the flow is weak; when they are narrowly spaced, the flow is stronger. The increase in winds on the chart results in a stronger Coriolis force *(CF)*, which balances a larger pressure gradient force *(PGF)*.



FOCUS ON AN ADVANCED TOPIC

A Mathematical Look at the Geostrophic Wind

We know from an earlier discussion that the geostrophic wind gives us a good approximation of the real wind above the level of friction, about 500 to 1000 m above the earth's surface. Above the friction layer, the winds tend to blow parallel to the isobars, or contours. We know that, for any given latitude, the speed of the geostrophic wind is proportional to the pressure gradient. This may be represented as

$$V_g \sim \frac{\Delta p}{d},$$

where Vg is the geostrophic wind and Δp is the pressure difference between two places some horizontal distance (d) apart. From this, we can see that the greater the pressure gradient, the stronger the geostrophic wind.

When we consider a unit mass of moving air, we must take into account the air density (mass per unit volume) expressed by the symbol ρ. The geostrophic wind is now directly proportional to the pressure gradient force; thus

$$V_g \sim \frac{1}{\rho}\frac{\Delta p}{d}.$$

We can see from this expression that, with the same pressure gradient (at the same latitude), the geostrophic wind will increase with increasing elevation because air density decreases with height.

In a previous section, we saw that the geostrophic wind represents a balance of forces between the Coriolis force and the pressure gradient force. Here, it should be noted that the Coriolis force (per unit mass) can be expressed as

$$\text{Coriolis force} = 2\Omega V \sin\phi,$$

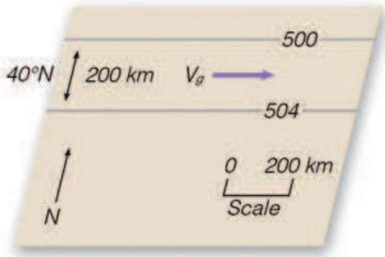


FIGURE 4 A portion of an upper-air chart for part of the Northern Hemisphere at an altitude of 5600 meters above sea level. The lines on the chart are isobars, where 500 equals 500 millibars. The air temperature is $-25°C$ and the air density is 0.70 kg/m^3.

where Ω is the earth's angular spin (a constant), V is the speed of the wind, and ϕ is the latitude. The $\sin\phi$ is a trigonometric function that takes into account the variation of the Coriolis force with latitude. At the equator (0°), $\sin\phi$ is 0; at 30° latitude, $\sin\phi$ is 0.5, and, at the poles (90°), $\sin\phi$ is 1.

This balance between the Coriolis force and the pressure gradient force can be written as

$$CF = PGF$$

$$2\Omega V_g \sin\phi = \frac{1}{\rho}\frac{\Delta p}{d}. \qquad (1)$$

Solving for V_g, the geostrophic wind, the equation becomes

$$V_g = \frac{1}{2\Omega\sin\phi\,\rho}\frac{Dp}{d}.$$

Customarily, the rotational (2Ω) and latitudinal ($\sin\phi$) factors are combined into a single value f, called the *Coriolis parameter*. Thus, we have the geostrophic wind equation written as

$$V_g \sim \frac{1}{f\rho}\frac{\Delta p}{d}. \qquad (2)$$

Suppose we compute the geostrophic wind for the example given in Fig. 4. Here the wind is blowing parallel to the isobars in the Northern Hemisphere at latitude 40°. The spacing between the isobars is 200 km and the pressure difference is 4 mb. The altitude is 5600 m above sea level, where the air temperature is $-25°C$ ($-13°F$) and the air density is 0.70 kg/m^3. First, we list our data and put them in the proper units, as

$$\Delta p = 4 \text{ mb} = 400 \text{ Newtons/m}^2$$
$$d = 200 \text{ km} = 2 \times 10^5 \text{ m}$$
$$\sin\phi = \sin(40°) = 0.64$$
$$\rho = 0.70 \text{ kg/m}^3$$
$$2\Omega = 14.6 \times 10^{-5} \text{ radian/sec.*}$$

When we use equation (1) to compute the geostrophic wind, we obtain

$$V_g = \frac{1}{2\Omega\sin\phi\rho}\frac{\Delta p}{d},$$

$$V_g = \frac{400}{14.6 \times 10^{-5} \times 0.64 \times 0.70 \times 2 \times 10^5},$$

$$V_g = 30.6 \text{ m/sec, or } 59.4 \text{ knots.}$$

*The rate of the earth's rotation (Ω) is 360° in one day, actually a sidereal day consisting of 23 hr, 56 min, 4 sec, or 86,164 seconds. This gives a rate of rotation of 4.18 $\times 10^{-3}$ degree per second. Most often, Ω is given in radians, where 2π radians equals 360° ($\pi = 3.14$). Therefore, the rate of the earth's rotation can be expressed as 2π radians/86,164 sec, or 7.29×10^{-5} radian/sec, and the constant 2Ω becomes 14.6×10^{-5} radian/sec.

View this concept in action on the Meteorology Resource Center at academic.cengage.com/login

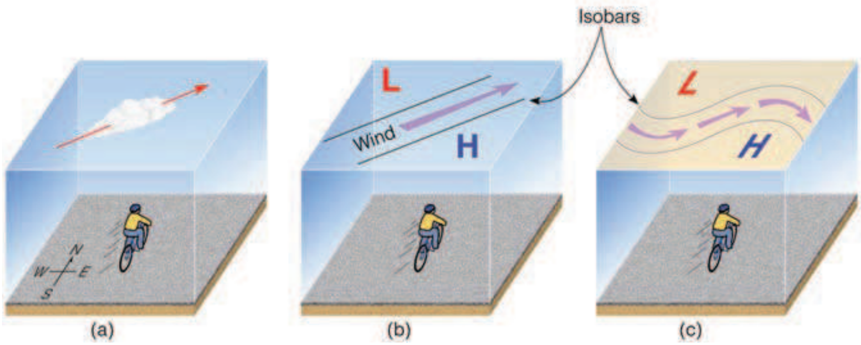
CURVED WINDS AROUND LOWS AND HIGHS ALOFT—GRADIENT WINDS Because lows are also known as cyclones, the counterclockwise flow of air around them is often called *cyclonic flow*. Likewise, the clockwise flow of air around a high, or anticyclone, is called *anticyclonic flow*. Look at the wind flow around the upper-level low (Northern Hemisphere) in Fig. 8.27. At first, it appears as though the wind is defying the Coriolis force by bending to the left as it moves counterclockwise around the system. Let's see why the wind blows in this manner.

FOCUS ON AN OBSERVATION

Estimating Wind Direction and Pressure Patterns Aloft by Watching Clouds

Both the wind direction and the orientation of the isobars aloft can be estimated by observing middle- and high-level clouds from the earth's surface. Suppose, for example, we are in the Northern Hemisphere watching clouds directly above us move from southwest to northeast at an elevation of about 3000 m or 10,000 ft (see Fig. 5a). This indicates that the geostrophic wind at this level is southwesterly. Looking downwind, the geostrophic wind blows parallel to the isobars with lower pressure on the left and higher pressure on the right. Thus, if we *stand with our backs to the direction from which the clouds are moving, lower pressure aloft will always be to our left and higher pressure to our right.* From this observation, we can draw a rough upper-level chart (see Fig. 5b), which shows isobars and wind direction for an elevation of approximately 10,000 ft.

The isobars aloft will not continue in a southwest-northeast direction indefinitely; rather, they will often bend into wavy patterns. We may carry our observation one step farther, then, by assuming a bending of the lines (Fig. 5c). Thus, with a southwesterly wind aloft, a trough of low pressure will be found to our west and a ridge of high pressure to our east. What would be the pressure pattern if the winds aloft were blowing *from* the northwest? Answer: A trough would be to the east and a ridge to the west.

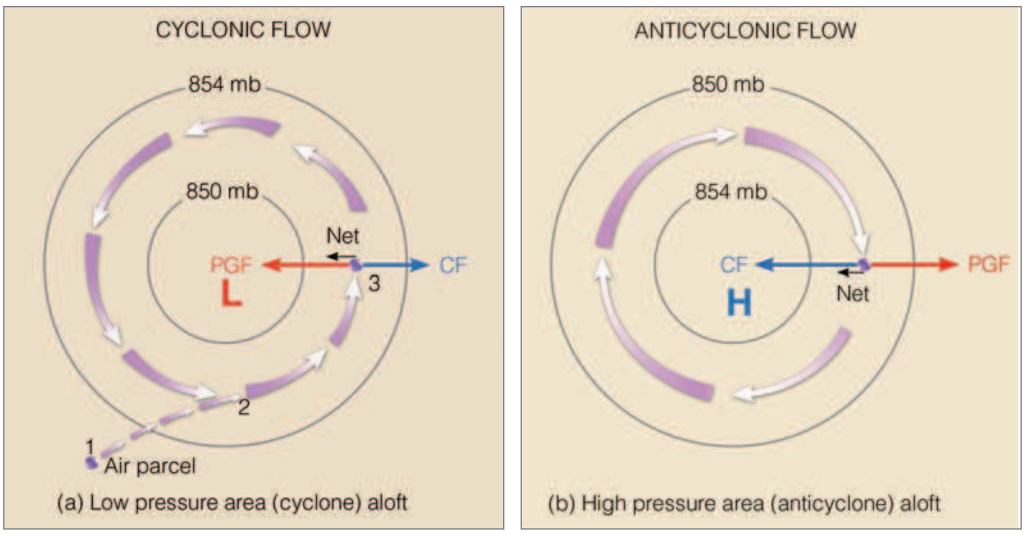


● FIGURE 5 This drawing of a simplified upper-level chart is based on cloud observations. Upper-level clouds moving from the southwest (a) indicate isobars and winds aloft (b). When extended horizontally, the upper-level chart appears as in (c), where a trough of low pressure is to the west and a ridge of high pressure is to the east.

Suppose we consider a parcel of air initially at rest at position 1 in Fig. 8.27a. The pressure gradient force accelerates the air inward toward the center of the low and the Coriolis force deflects the moving air to its right, until the air is moving parallel to the isobars at position 2. If the wind were geostrophic, at position 3 the air would move northward parallel to straight-line isobars at a constant speed. The wind is blowing at a constant speed, but parallel to curved isobars. A wind that blows at a constant speed parallel to *curved isobars* above the level of frictional influence is termed a **gradient wind.**

Earlier in this chapter we learned that an object accelerates when there is a change in its speed or direction (or both). Therefore, the gradient wind blowing *around* the low-pressure center is constantly accelerating because it is con-

● FIGURE 8.27 Winds and related forces around areas of low and high pressure above the friction level in the Northern Hemisphere. Notice that the pressure gradient force *(PGF)* is in red, while the Coriolis force *(CF)* is in blue.



stantly changing direction. This acceleration, called the **centripetal acceleration,** is directed at right angles to the wind, inward toward the low center.

Remember from Newton's second law that, if an object is accelerating, there must be a *net force* acting on it. In this case, the net force acting on the wind must be directed toward the center of the low, so that the air will keep moving in a circular path. This inward-directed force is called the **centripetal force** (*centri:* center; *petal:* to push toward). The magnitude of the centripetal force is related to the wind velocity *(V)* and the radius of the wind's path *(r)* by the formula

$$\text{Centripetal force} = \frac{V^2}{r}.$$

Where wind speeds are light and there is little curvature (large radius), the centripetal force is weak and, compared to other forces, may be considered insignificant. However, where the wind is strong and blows in a tight curve (small radius), as in the case of tornadoes and tropical hurricanes, the centripetal force is large and becomes quite important.

The centripetal force results from an imbalance between the Coriolis force and the pressure gradient force.* Again, look closely at position 3 (Fig. 8.27a) and observe that the inward-direction pressure gradient force *(PGF)* is greater than the outward-directed Coriolis force *(CF).* The difference between these two forces—the net force—is the inward-directed centripetal force. In Fig. 8.27b, the wind blows clockwise around the center of the high. The spacing of the isobars tells us that the magnitude of the *PGF* is the same as in Fig. 8.27a. However, to keep the wind blowing in a circle, the inward-directed Coriolis force must now be greater in magnitude than the outward-directed pressure gradient force, so that the centripetal force (again, the net force) is directed inward.

The greater Coriolis force around the high results in an interesting observation. Because the Coriolis force (at any given latitude) can increase only when the wind speed increases, we can see that for the same pressure gradient (the same spacing of the isobars), the winds around a high-pressure area (or a ridge) must be greater than the winds around a low-pressure area (or a trough). Normally, however, the winds blow much faster around an area of low pressure (a cyclonic storm) than they do around an area of high pressure because the isobars around the low are usually spaced much closer together, resulting in a much stronger pressure gradient.

In the Southern Hemisphere, the pressure gradient force starts the air moving, and the Coriolis force deflects the moving air to the *left,* thereby causing the wind to blow *clockwise around lows* and *counterclockwise around highs.* ● Figure 8.28 shows a satellite image of clouds and wind flow (dark arrows) around a low-pressure area in the Northern Hemisphere (8.28a) and in the Southern Hemisphere (8.28b).

Near the equator, where the Coriolis force is minimum, winds may blow around intense tropical storms with the centripetal force being almost as large as the pressure gradient force. In this type of flow, the Coriolis force is considered negligible, and the wind is called *cyclostrophic.*

So far we have seen how winds blow in theory, but how do they appear on an actual map?

WINDS ON UPPER-LEVEL CHARTS

On the upper-level 500-mb map (● Figure 8.29), notice that, as we would expect, the winds tend to parallel the contour lines in a wavy west-to-east direction. Notice also that the contour lines tend to decrease in elevation from south to north. This situation occurs because the air at this level is warmer to the south and colder to the north. On the map, where horizontal temperature contrasts are large there is also a large height gradient—the contour lines are close together and the winds are strong. Where the horizontal temperature contrasts are small, there is a small height gradient—the contour lines are spaced farther apart and the winds are weaker. In general, on maps such as this we find stronger north-to-south temperature contrasts in winter than in summer, which is why the winds aloft are usually stronger in winter.

In Fig. 8.29, the wind is geostrophic where it blows in a straight path parallel to evenly spaced lines; it is gradient where it blows parallel to curved contour lines. Where the wind flows in large, looping meanders, following a more or less north-south trajectory (such as along the west coast of North America), the wind-flow pattern is called **meridional.** Where the winds are blowing in a west-to-east direction (such as over the eastern third of the United States), the flow is termed **zonal.**

Because the winds aloft in middle and high latitudes generally blow from west to east, planes flying in this direction have a beneficial tail wind, which explains why a flight from San Francisco to New York City takes about thirty minutes less than the return flight. If the flow aloft is zonal, clouds, storms, and surface anticyclones tend to move more rapidly from west to east. However, where the flow aloft is meridional, as we will see in Chapter 12 surface storms tend to move more slowly, often intensifying into major storm systems.

We know that the winds aloft in the middle latitudes of the Northern Hemisphere tend to blow in a west-to-east pattern. Does this mean that the winds aloft in the Southern Hemisphere blow from east-to-west? If you are unsure of the answer, read the Focus section on p. 215.

*In some cases, it is more convenient to express the centripetal force (and the centripetal acceleration) as the *centrifugal force,* an apparent force that is equal in magnitude to the centripetal force, but directed outward from the center of rotation. The gradient wind is then described as a balance of forces between the centrifugal force

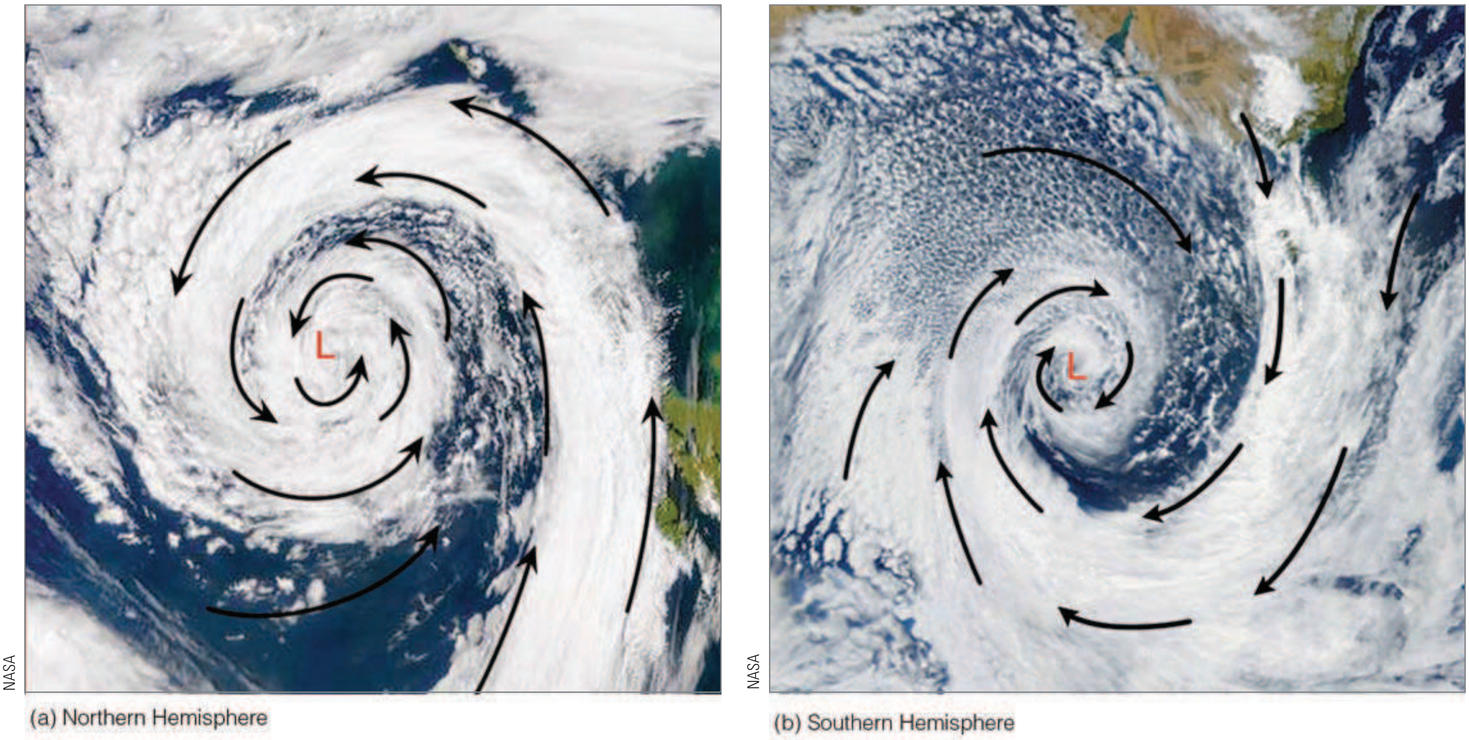
$$\frac{V^2}{r},$$

the pressure gradient force

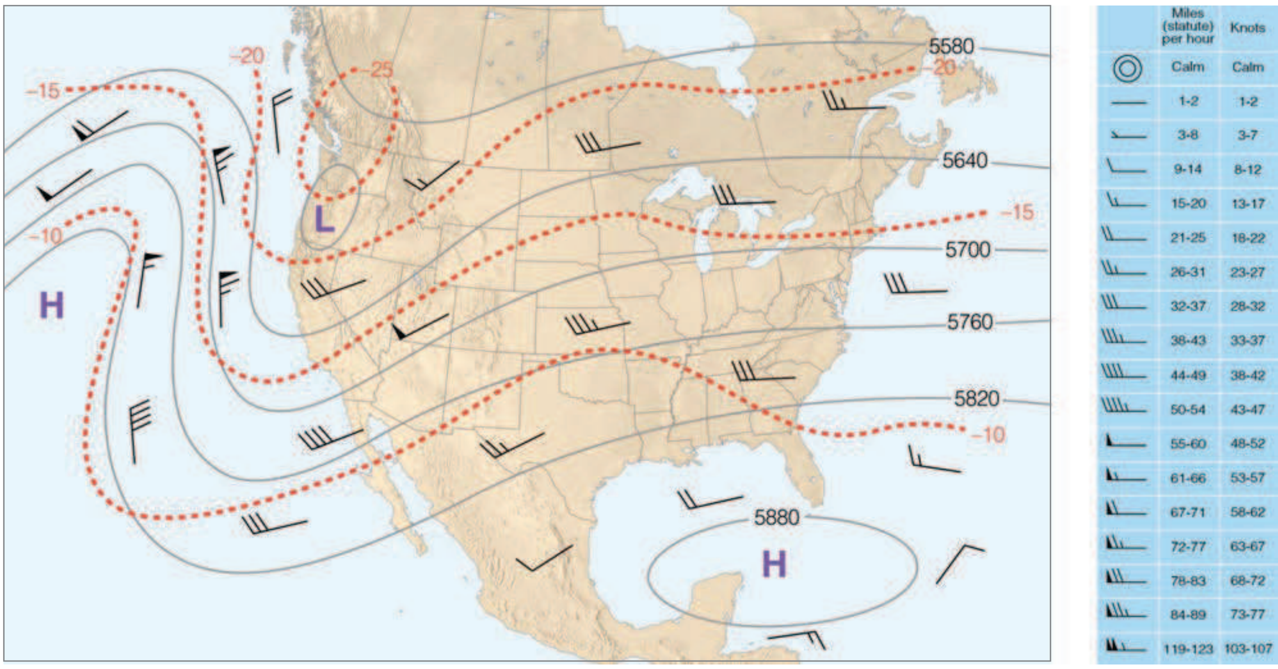
$$\frac{1}{\rho}\frac{\Delta p}{d},$$

and the Coriolis force $2\Omega V \sin \phi$. Under these conditions, the *gradient wind equation* for a unit mass of air is expressed as

$$\frac{V^2}{r} + \frac{1}{\rho}\frac{\Delta p}{d} + 2\Omega V \sin \phi = 0.$$



● FIGURE 8.28 Clouds and related wind-flow patterns (black arrows) around low-pressure areas. (a) In the Northern Hemisphere, winds blow counterclockwise around an area of low pressure. (b) In the Southern Hemisphere, winds blow clockwise around an area of low pressure.



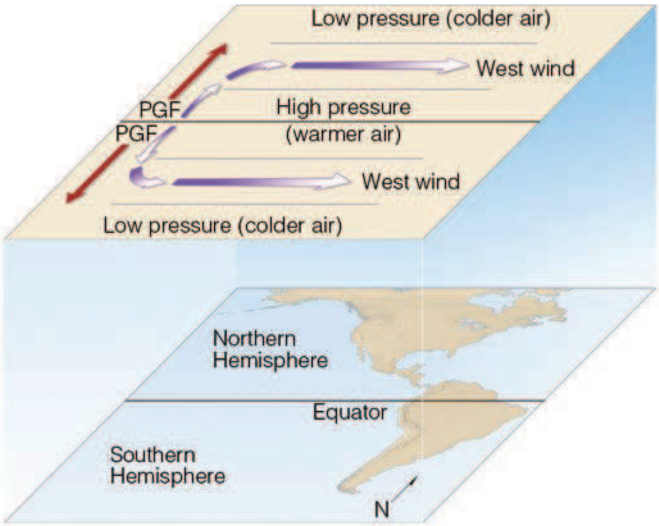
● FIGURE 8.29 An upper-level 500-mb map showing wind direction, as indicated by lines that parallel the wind. Wind speeds are indicated by barbs and flags. (See the blue insert.) Solid gray lines are contours in meters above sea level. Dashed red lines are isotherms in °C.

FOCUS ON AN OBSERVATION

Winds Aloft in the Southern Hemisphere

In the Southern Hemisphere, just as in the Northern Hemisphere, the winds aloft blow because of horizontal differences in pressure. The pressure differences, in turn, are due to variations in temperature. Recall from an earlier discussion of pressure that warm air aloft is associated with high pressure and cold air aloft with low pressure. Look at Fig. 6. It shows an upper-level chart that extends from the Northern Hemisphere into the Southern Hemisphere. Over the equator, where the air is warmer, the pressure aloft is higher. North and south of the equator, where the air is colder, the pressure aloft is lower.

Let's assume, to begin with, that there is no wind on the chart. In the Northern Hemisphere, the pressure gradient force directed northward starts the air moving toward lower pressure. Once the air is set in motion, the Coriolis force bends it to the right until it is a *west wind,* blowing parallel to the isobars. In the Southern Hemisphere, the pressure gradient force directed southward starts the air moving south. But notice that the Coriolis force in the Southern Hemisphere bends the moving air to its *left,* until the wind is blowing parallel to the isobars *from the west.* Hence, in the middle and high latitudes of both hemispheres, we generally find westerly winds aloft.



● FIGURE 6
Upper-level chart that extends over the Northern and Southern Hemispheres. Solid gray lines on the chart are isobars.

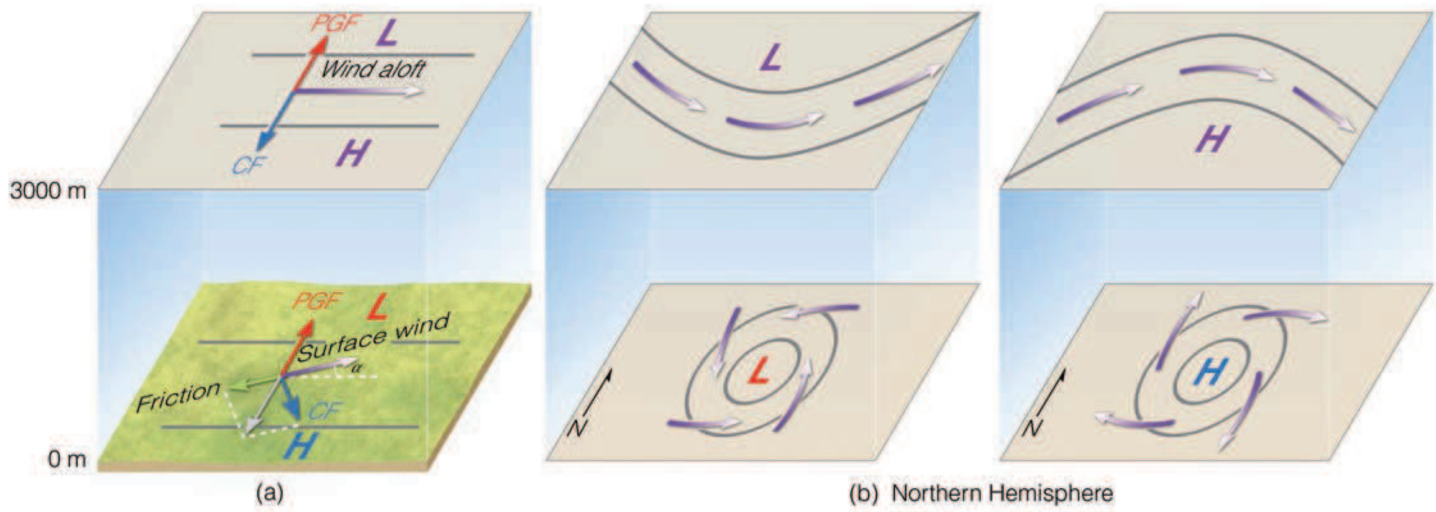
Take a minute and look back at Fig. 8.20 on p. 207. Observe that the winds on this surface map tend to cross the isobars, blowing from higher pressure toward lower pressure. Observe also that along the green line, the tightly packed isobars are producing a steady surface wind of 40 knots. However, this same pressure gradient (with the same air temperature) would, on an upper-level chart, produce a much stronger wind. Why do surface winds normally cross the isobars and why do they blow more slowly than the winds aloft? The answer to both of these questions is *friction.*

SURFACE WINDS The frictional drag of the ground slows the wind down. Because the effect of friction decreases as we move away from the earth's surface, wind speeds tend to increase with height above the ground. The atmospheric layer that is influenced by friction, called the **friction layer** (or *planetary boundary layer*), usually extends upward to an altitude near 1000 m (3300 ft) above the surface, but this altitude may vary due to strong winds or irregular terrain. (We will examine the planetary boundary layer winds more thoroughly in Chapter 9.)

In ● Fig. 8.30a, the wind aloft is blowing at a level above the frictional influence of the ground. At this level, the wind is approximately geostrophic and blows parallel to the isobars, with the pressure gradient force *(PGF)* on its left balanced by the Coriolis force *(CF)* on its right. At the earth's surface, the same pressure gradient will not produce the same wind speed, and the wind will not blow in the same direction.

Near the surface, *friction reduces the wind speed, which in turn reduces the Coriolis force.* Consequently, the weaker Coriolis force no longer balances the pressure gradient force, and the wind blows across the isobars toward lower pressure. The angle (α) at which the wind crosses the isobars varies, but averages about 30°.* As we can see in Fig. 8.30a, at the surface the pressure gradient force is now balanced by the sum of the frictional force and the Coriolis force. Therefore, in the Northern Hemisphere, we find surface winds blowing counterclockwise and *into* a low; they flow clockwise and *out* of a high (see Fig. 8.30b). In the Southern Hemisphere, winds blow clockwise and inward around surface lows; counter-

*The angle at which the wind crosses the isobars to a large degree depends upon the roughness of the terrain. Everything else being equal, the rougher the surface, the larger the angle. Over hilly land, the angle might average between 35° and 40°, while over an open body of relatively smooth water it may average between 10° and 15°. Taking into account all types of surfaces, the average is near 30°. This angle also depends on the wind speed. Typically, the angle is smallest for high winds and largest for gentle breezes. As we move upward through the friction layer, the wind becomes more and more parallel to the isobars.



● FIGURE 8.30 (a) The effect of surface friction is to slow down the wind so that, near the ground, the wind crosses the isobars and blows toward lower pressure. (b) This phenomenon at the surface produces an inflow of air around a low and an outflow of air around a high. Aloft, away from the influence of friction, the winds blow parallel to the lines, usually in a wavy west-to-east pattern.

clockwise and outward around surface highs (see ● Fig. 8.31). ● Figure 8.32 illustrates a surface weather map and the general wind-flow pattern on a particular day in South America.

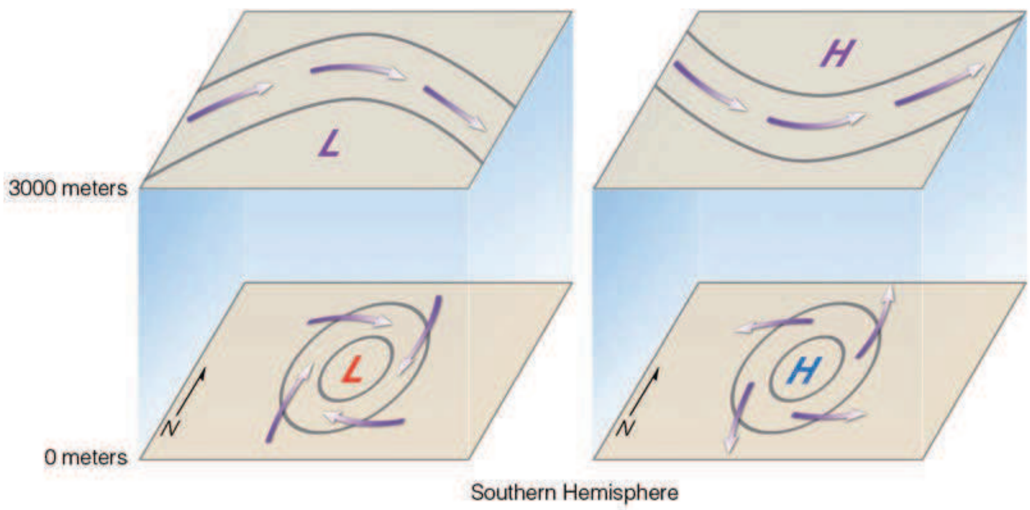
We know that, because of friction, surface winds move more slowly than do the winds aloft with the same pressure gradient. Surface winds also blow across the isobars toward lower pressure. The angle at which the winds cross the isobars depends upon surface friction, wind speed, and the height above the surface. Aloft, however, the winds blow parallel to contour lines, with lower pressure (in the Northern Hemisphere) to their left. Consequently, because of this fact, if you (in the Northern Hemisphere) stand with the wind aloft to your back, lower pressure will be to your left and higher pressure to your right (see ● Fig. 8.33a). The same rule applies to the surface wind, but with a slight modification due to the fact that here the wind crosses the isobars. Look at Fig. 8.33b and notice that, at the surface, *if you stand with your back to the wind, then turn clockwise about 30°, the center of lowest pressure will be to your left.** This relationship between wind and pressure is often called **Buys-Ballot's law,** after the Dutch meteorologist Christoph Buys-Ballot (1817–1890), who formulated it.

Winds and Vertical Air Motions

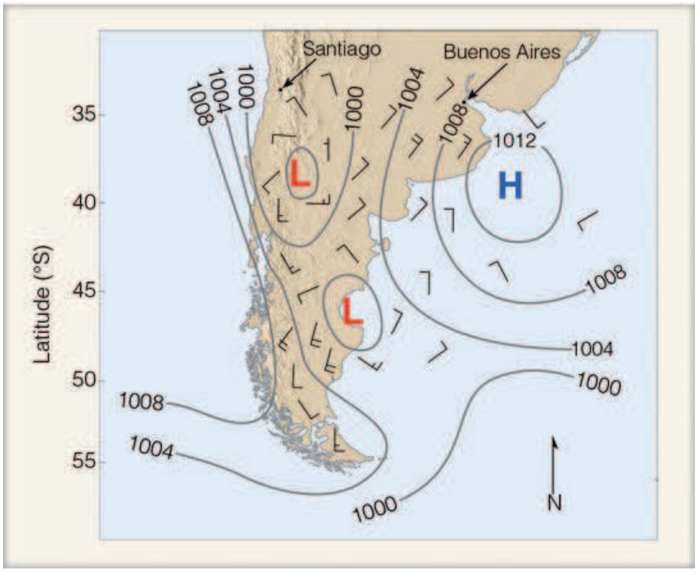
Up to this point, we have seen that surface winds blow in toward the center of low pressure and outward away from the center of high pressure. Notice in ● Fig. 8.34 that as air moves inward toward the center of low pressure, it must go somewhere. Since this converging air cannot go into the ground, it slowly rises. Above the surface low (at about 6 km or so), the air begins to diverge (spread apart).

As long as the upper-level diverging air balances the converging surface air, the central pressure in the surface low

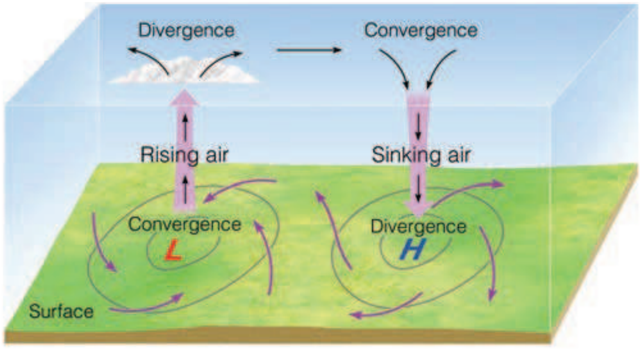
*In the Southern Hemisphere, stand with your back to the wind, then turn counterclockwise about 30°—the center of lowest pressure will then be to your right.



● FIGURE 8.31 Winds around an area of (a) low pressure and (b) high pressure in the Southern Hemisphere.



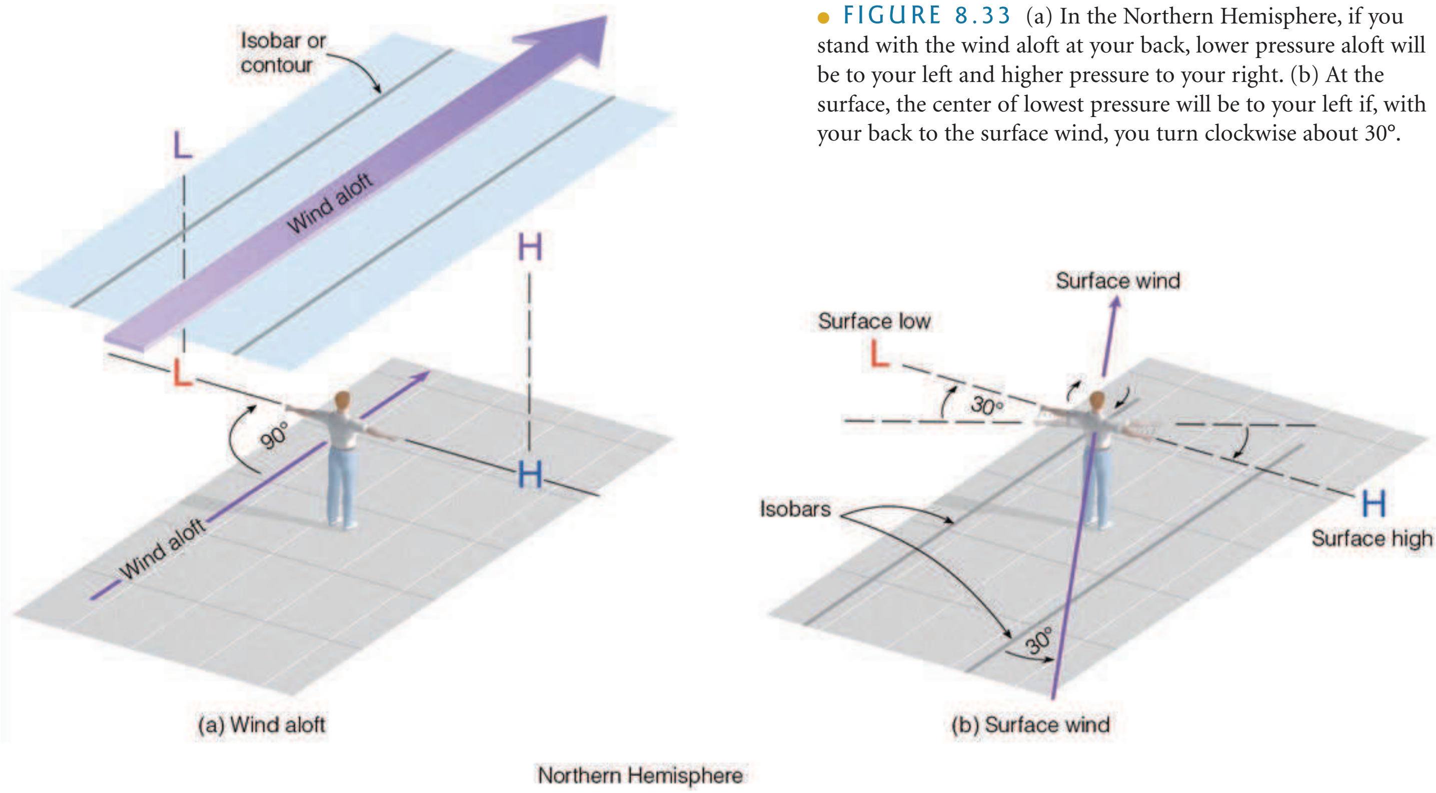
● FIGURE 8.32 Surface weather map showing isobars and winds on a day in December in South America.



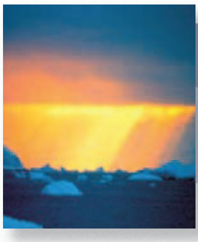
● FIGURE 8.34 Winds and air motions associated with surface highs and lows in the Northern Hemisphere.

does not change. However, the surface pressure *will change* if upper-level divergence and surface convergence are not in balance. For example, as we saw earlier in this chapter (when we examined the air pressure above two cities), the surface pressure will change if the mass of air above the surface changes. Consequently, if upper-level divergence exceeds surface convergence (that is, more air is removed at the top than is taken in at the surface), the air pressure at the center of the surface low will decrease, and isobars around the low will become more tightly packed. This situation increases the pressure gradient (and, hence, the pressure gradient force), which, in turn, increases the surface winds.

Surface winds move outward (diverge), away from the center of a high-pressure area. To replace this laterally spreading air, the air aloft converges and slowly descends as shown in Fig. 8.34. Again, as long as upper-level converging air balances surface diverging air, the air pressure in the center of the high will not change. (Convergence and divergence of air are so important to the development or weakening of surface pressure systems that we will examine this topic again when



● FIGURE 8.33 (a) In the Northern Hemisphere, if you stand with the wind aloft at your back, lower pressure aloft will be to your left and higher pressure to your right. (b) At the surface, the center of lowest pressure will be to your left if, with your back to the surface wind, you turn clockwise about 30°.



FOCUS ON AN ADVANCED TOPIC

The Hydrostatic Equation

Air is in hydrostatic equilibrium when the upward-directed pressure gradient force is exactly balanced by the downward force of gravity. Figure 7 shows air in hydrostatic equilibrium. Since there is no net vertical force acting on the air, there is no net vertical acceleration, and the sum of the forces is equal to zero, all of which is represented by

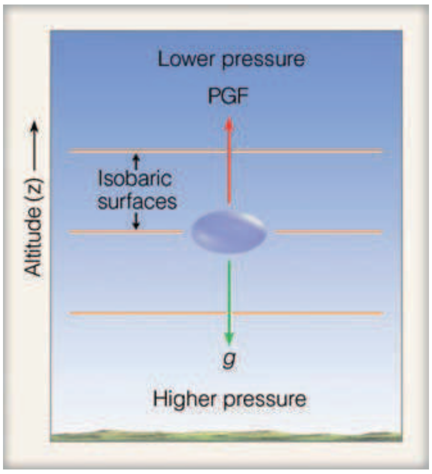
$$PGF_{\text{vertical}} + g = 0$$
$$\frac{1}{\rho}\frac{\Delta p}{\Delta z} + g = 0,$$

where ρ is the air density, Δp is the decrease in pressure along a small change in height (Δz), and g is the force of gravity. This expression is usually given as

$$\frac{\Delta p}{\Delta z} = -\rho g.$$

This equation is called the *hydrostatic equation.* The hydrostatic equation tells us that the rate at which the pressure decreases with height is equal to the air density times the acceleration of gravity (where ρg is actually the force of gravity per unit volume). The minus sign indicates that, as the air pressure decreases, the height increases. When the hydrostatic equation is given as

$$\Delta p = -\rho g\,\Delta z,$$



● FIGURE 7 When the vertical pressure gradient force (PGF) is in balance with the force of gravity (g), the air is in hydrostatic equilibrium.

it tells us something important about the atmosphere that we learned earlier: The air pressure decreases more rapidly with height in cold (more-dense) air than it does in warm (less-dense) air. In addition, we can use the hydrostatic equation to determine how rapidly the air pressure decreases with increasing height above the surface. For example, suppose at the surface a 1000 meter-thick layer of air (under standard conditions) has an average density of 1.1 kg/m³ and an acceleration of gravity of 9.8 m/sec². Therefore, we have

$$\rho = 1.1\ \text{kg/m}^3$$
$$g = 9.8\ \text{m/sec}^2$$
$$\Delta z = 1000\ \text{m}$$

(This value is the height difference from the surface [0 m] to an altitude of 1000 m.)

Using the hydrostatic equation to compute Δp, the difference in pressure in a 1000-meter-thick layer of air, we obtain

$$\Delta p = \rho g\,\Delta z$$
$$\Delta p = (1.1)\,(9.8)\,(1000)$$
$$\Delta p = 10{,}780\ \text{Newtons/m}^2.$$

Since $1\ \text{mb} = 100\ \text{Newtons/m}^2$,

$$\Delta p = 108\ \text{mb}.$$

Hence, air pressure decreases by about 108 mb in a standard 1000-meter layer of air near the surface. This closely approximates the pressure change of 10 mb per 100 meters we used in converting station pressure to sea-level pressure earlier in this chapter.

we look more closely at the vertical structure of pressure systems in Chapter 12.)

The rate at which air rises above a low or descends above a high is small compared to the horizontal winds that spiral about these systems. Generally, the vertical motions are usually only about several centimeters per second, or about 1.5 km (or 1 mi) per day.

Earlier in this chapter we learned that air moves in response to pressure differences. Because air pressure decreases rapidly with increasing height above the surface, there is always a strong pressure gradient force directed upward, much stronger than in the horizontal. Why, then, doesn't the air rush off into space?

Air does not rush off into space because the upward-directed pressure gradient force is nearly always exactly balanced by the downward force of gravity. When these two forces are in exact balance, the air is said to be in **hydrostatic equilibrium.** When air is in hydrostatic equilibrium, there is no net vertical force acting on it, and so there is no net vertical acceleration. Most of the time, the atmosphere approximates hydrostatic balance, even when air slowly rises or descends at a constant speed. However, this balance does not exist in violent thunderstorms and tornadoes, where the air shows appreciable vertical acceleration. But these occur over relatively small vertical distances, considering the total vertical extent of the atmosphere. (A more mathematical look at hydrostatic equilibrium, expressed by the *hydrostatic equation,* is given in the Focus section above.)

SUMMARY

This chapter gives us a broad view of how and why the wind blows. We examined constant pressure charts and found that low heights correspond to low pressure and high heights to high pressure. In regions where the air aloft is cold, the air pressure is normally lower than average; where the air aloft is warm, the air pressure is normally higher than average. Where horizontal variations in temperature exist, there is a corresponding horizontal change in pressure. The difference in pressure establishes a force, the pressure gradient force, which starts the air moving from higher toward lower pressure.

Once the air is set in motion, the Coriolis force bends the moving air to the right of its intended path in the Northern Hemisphere and to the left in the Southern Hemisphere. Above the level of surface friction, the wind is bent enough so that it blows nearly parallel to the isobars, or contours. Where the wind blows in a straight-line path, and a balance exists between the pressure gradient force and the Coriolis force, the wind is termed geostrophic. Where the wind blows parallel to curved isobars (or contours), the centripetal acceleration becomes important, and the wind is called a gradient wind. When the wind-flow pattern aloft is west-to-east, the flow is called *zonal;* where the wind flow aloft is more north-south, the flow is called *meridional.*

The interaction of the forces causes the wind in the Northern Hemisphere to blow clockwise around regions of high pressure and counterclockwise around areas of low pressure. In the Southern Hemisphere, the wind blows counterclockwise around highs and clockwise around lows. The effect of surface friction is to slow down the wind. This causes the surface air to blow across the isobars from higher pressure toward lower pressure. Consequently, in both hemispheres, surface winds blow outward, away from the center of a high, and inward, toward the center of a low.

When the upward-directed pressure gradient force is in balance with the downward force of gravity, the air is in hydrostatic equilibrium. Since there is no net vertical force acting on the air, it does not rush off into space.

KEY TERMS

The following terms are listed (with page numbers) in the order they appear in the text. Define each. Doing so will aid you in reviewing the material covered in this chapter.

QUESTIONS FOR REVIEW

1. Why does air pressure decrease with height more rapidly in cold air than in warm air?
2. What can cause the air pressure to change at the bottom of a column of air?
3. What is considered standard sea-level atmospheric pressure in millibars? In inches of mercury? In hectopascals?
4. How does an aneroid barometer differ from a mercury barometer?
5. How does sea-level pressure differ from station pressure? Can the two ever be the same? Explain.
6. On an upper-level chart, is cold air aloft generally associated with low or high pressure? What about warm air aloft?
7. What do Newton's first and second laws of motion tell us?
8. Explain why, in the Northern Hemisphere, the average height of contour lines on an upper-level isobaric chart tend to decrease northward.
9. What is the force that initially sets the air in motion?
10. What does the Coriolis force do to moving air (a) in the Northern Hemisphere? (b) in the Southern Hemisphere?
11. Explain how each of the following influences the Coriolis force: (a) rotation of the earth; (b) wind speed; (c) latitude.
12. How does a steep (or strong) pressure gradient appear on a weather map?
13. Explain why on a map, closely spaced isobars (or contours) indicate strong winds, and widely spaced isobars (or contours) indicate weak winds.
14. What is a geostrophic wind? Why would you *not* expect to observe a geostrophic wind at the equator?
15. Why do upper-level winds in the middle latitudes of both hemispheres generally blow from the west?
16. Describe how the wind blows around highs and lows aloft and near the surface (a) in the Northern Hemisphere and (b) in the Southern Hemisphere.
17. What are the forces that affect the horizontal movement of air?
18. What factors influence the angle at which surface winds cross the isobars?
19. Describe the type of vertical air motions associated with surface high- and low-pressure areas.
20. Since there is always an upward-directed pressure gradient force, why doesn't the air rush off into space?