THE ATMOSPHERE IN MOTION:
AIR PRESSURE, FORCES, AND WINDS

December 19, 1930, 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 able to predict that the wind would ruin the advertising scheme.
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 weight 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. But there are other important concepts to consider. For example, air pressure, air density (the mass of air in a given volume), and air temperature 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. 212.

To help eliminate some of the complexities of the atmosphere, scientists construct models. Figure 9.1 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, and (2) that the width of the column does not change with height.

Suppose we somehow force more air into the column in Fig. 9.1. 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 mass 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. With the same assumptions for our model in Fig. 9.1, let's look at Fig. 9.2.

**FIGURE 9.1**
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.

Suppose the two air columns in Fig. 9.2a are located at the same elevation and have identical surface air pressures. 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. 9.2b).

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 surface pressure does not vary and the total number of molecules above each city must remain the same. 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 column of air above city 1 and a warm taller 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 elevation in the cold column of air. In the cold air above city 1 (Fig. 9.2b), 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. 9.2c, move up the warm column until you come to the letter H. Now move up the cold 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 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 cool air aloft is associated with low atmospheric pressure.
In Fig. 9.2c, 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.

In summary, heating or cooling a column of air can establish horizontal variations in pressure that cause the air to move. The net accumulation of air above the surface causes the surface air pressure to rise, whereas a decrease in the amount of air above the surface causes the surface air pressure to fall.

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 accompanied 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. 9.3). Maximum pressures occur around 10:00 A.M. and 10:00 P.M., minimum near...
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

\[ p = T \times \rho \]

where, of course, \( p \) is pressure, \( T \) is temperature, and \( \rho \) (the Greek letter rho, pronounced "row") represents air density. The line \( \sim \) 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

\[ p \sim \rho \] (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 storms) to form, 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.

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.

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. 9.4 compares pressure readings in millibars and in inches of mercury.

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 preferred...
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

\[(\text{Constant pressure}) \times \text{constant} = T \times p.\]

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 mb. 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

\[p = T \times p \times C.\]

With the pressure \(p\) in millibars (mb), the temperature \(T\) in Kelvin, and the density \(p\) in kilograms per cubic meter \((\text{kg/m}^3)\), 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 \(\text{kg/m}^3\), the average air temperature becomes

\[p = T \times p \times C\]
\[500 = T \times 0.690 \times 2.87\]
\[0.590 \times 2.87 = T\]
\[252.5 \text{ K} = T\]

To convert Kelvin into degrees Celsius, we subtract 273 from the Kelvin

\[\text{WEATHER WATCH}\]

Mercury is 13.6 times more dense than water. Consequently, a water barometer resting on the ground near sea level would have to be read from a ladder over 10.3 m (34 ft) tall.

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.
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. 9.6).

Notice that the aneroid barometer often has descriptive weather-related words printed above specific pressure values. These adjectives 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 phenomenon 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.

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. 9.7).

**Pressure Readings** The seemingly simple task of reading the height of the mercury column to obtain the air pressure is actually not at 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 9.8a 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 \( \frac{1}{2} \) 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 an altitude of 0 meters, which is referred to as mean sea level (MSL)—the level representing the average surface of the ocean. The adjusted pressure 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 by about 10 mb for every 100 m increase in elevation (about 1 in. of mercury for each 1000-ft rise) in an atmosphere where the air temperature decreases at the standard lapse rate of 6.5°C/1000 m. Therefore, we generally add 10 mb for every 100 m of altitude to obtain sea-level pressure.

Because a standard atmosphere seldom exists, this value (10 mb per 100 m) only approximates actual conditions. For example, as we saw earlier in this chapter, atmospheric pressure decreases more rapidly with height in cold (more-dense) air than it does in warm (less-dense) air. Hence, in cold air, the vertical rate of pressure change is typically greater than 10 mb per 100 m, whereas in warm air, it is less. Moreover, since the vertical lapse rate changes slightly from day to day, temperature and moisture corrections made for "average" conditions may introduce additional errors. However, if we simplify matters and assume a standard lapse rate, we can correct each of the station pressures in Fig. 9.8a to sea level simply by adding 10 mb per 100 m to each station pressure. When we plot these corrected values on the sea-level pressure chart (Fig. 9.8b), we are able to see the horizontal variation in sea-level pressure. Notice that the pressure varies from higher pressure on the left side of the map to lower pressure on the right—something impossible to see from the uncorrected station pressures we initially examined in Fig. 9.8a.

When more pressure data are added (see Fig. 9.8c), the chart can be analyzed and the pressure pattern visualized. 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. 9.8c) 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. 9.9a 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 wiggles 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 3098 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 9.9b 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. 9.10).

Another type of chart is commonly used in studying the weather, namely, the constant pressure (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 the light blue shaded area inside the air column in Fig. 9.11 represents tightly packed air molecules from the surface up to 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 gray 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.

*An interval of 2 mb would put the lines too close together, and an 8-mb interval would spread them too far apart.
FIGURE 9.9
(a) Sea-level isobars drawn so that each observation is taken into account. Not all observations are plotted. (b) Sea-level isobars after smoothing.

FIGURE 9.10
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 9.11
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.
If the air temperature should change in any portion of the column, the air density and pressure would change along with it (see Fig. 9.12). Note 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. Consequently, the cold, more dense air column is shorter, while the warm, less dense air column is higher.

In Fig. 9.12, 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 air the 500-mb pressure surface is found at a higher (than average) level, while in the colder 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. 9.12 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 510 mb. Therefore, we can conclude that high heights on a constant pressure chart correspond to higher-than-normal pressures at any given altitude, and low heights on a constant pressure chart correspond to lower-than-normal pressures.

The variations in height of the constant pressure surface in Fig. 9.12 are shown in Fig. 9.13. 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 in the manner of isobars, as contour lines of low height represent a region of lower pressure and contour lines of high height represent a region of higher pressure.

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.
TABLE 9.1 Common Isobaric Charts and Their Approximate Elevation above Sea Level

<table>
<thead>
<tr>
<th>ISOBARIC SURFACE</th>
<th>APPROXIMATE ELEVATION</th>
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<tbody>
<tr>
<td>(MB) CHARTS</td>
<td>(m)</td>
</tr>
<tr>
<td>1000</td>
<td>120</td>
</tr>
<tr>
<td>850</td>
<td>1,460</td>
</tr>
<tr>
<td>700</td>
<td>3,000</td>
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<td>500</td>
<td>5,600</td>
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<td>300</td>
<td>9,180</td>
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<td>200</td>
<td>11,800</td>
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<tr>
<td>100</td>
<td>16,200</td>
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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. 9.14, we can see how the wavy contours on the map relate to the changes in altitude of the pressure surface.

Although we have examined only the 500-mb chart, other isobaric charts are commonly used. Table 9.1 lists these charts and their approximate heights above sea level.

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, Lock Out Below," p. 220.)

Figure 9.15a 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

(a) Surface map

(b) 500-millibar map

FIGURE 9.15
(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. (b) The 500-mb map for the same day as the surface map. Solid dark lines on the map are contour lines in meters above sea level. Dashed red lines are isotherms in °C. Arrows show wind direction.
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°F/1000 ft (3.6°C/1000 ft). Since the air temperature seldom, if ever, decreases at 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 heavy dashed 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. This event is especially likely close to the ground (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 now 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.

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.
cyclones because they form in the 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 9.15b shows an upper-air chart (a 500-mb map) for the same day as the surface map in Fig. 9.15a. The solid dark 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 m. 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. 9.15a, the winds on the 500-mb chart tend to flow parallel to the contour lines in a more or less west-to-east direction. Why do the wind tend to cross the isobars on a surface map, yet blow parallel to the contour lines (or isobars) on an upper-sir 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 it. 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. 9.16, 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

![Figure 9.16](image)

*Velocity specifies both the speed of an object and its direction of motion.
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 9.17 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

\[
\text{Pressure gradient} = \frac{\text{difference in pressure}}{\text{distance}}
\]

If we let the symbol delta (\(\Delta\)) mean “a change in,” we can simplify the expression and write the pressure gradient as

\[
PG = \frac{\Delta p}{d},
\]

where \(\Delta p\) is the pressure difference between two places some horizontal distance \(d\) apart. In Fig. 9.17 the pressure gradient between points 1 and 2 is 4 mb per 100 km.

Suppose the pressure in Fig. 9.17 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.

Notice in Fig. 9.17 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 9.18 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 chart 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. These rules also apply to surface weather maps. An example of a steep pressure gradient producing strong winds is given in Fig. 9.19. Notice that the tightly packed isobars along the green line are producing a steep pressure gradient of 32 mb per 500 km and 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. 9.20, 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 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. 9.20, 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...
The atmosphere is in motion, and this motion can affect the weather. The Coriolis effect, due to the rotation of the earth, causes winds to deflect to the right in the Northern Hemisphere. Surface weather maps, like the one in Figure 9.19, show the location of weather systems and the direction of the winds.

The Coriolis effect is important in understanding how weather systems move and how winds are deflected. For example, high-pressure systems move towards the east, while low-pressure systems move towards the west. This is why the weather map in Figure 9.19 shows a high-pressure system moving eastward.

Weather watch: The deep, low-pressure area illustrated in Fig. 9.19 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 and in Michigan's Mackinac Island. 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.

From above the South Pole, 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 9.22 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 deflection. 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. 9.22.

Imagine in Fig. 9.22 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. 9.22).
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 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.

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*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, \( \Omega \) is the earth’s angular rate of spin (a constant), \( V \) is the speed of the object, and \( \phi \) is the latitude.

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**Figure 9.22**

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.
WEATHER WATCH

If the Coriolis force acts on all moving objects, why doesn’t it pull your moving car to the right? Actually, when you drive along a highway (at the speed limit), the Coriolis force would “pull” your car to the right about 460 m (1500 ft) for every 160 km (100 mi) you travel, if it were not for the friction between your tires and the road surface.

...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 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. 9.24, 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. 226.)

In Fig. 9.25, 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

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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. 9.23, which shows a map of the Northern Hemisphere, above the earth’s frictional influence,* with horizontal pressure variations at an altitude of about 1 km. The evenly spaced isobars indicate a constant pressure gradient force (PGF) directed from south toward north as indicated by the arrow at the left. Why, then, does the map show a west wind? 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 (geo: earth; strophic: turning) wind. Notice that the

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*We will see later that 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 9.23

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.
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:

\[ V_g = \frac{\Delta p}{d} \]

where \( V_g \) is the geostrophic wind and \( \Delta p \) 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 \( \rho \). 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. Hence, it should be noted that the Coriolis force can be expressed as

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

where \( \Omega \) is the earth's angular spin (a constant), \( V \) is the speed of the wind, and \( \phi \) is the latitude. The \( \sin \phi \) is a trigonometric function that takes into account the variation of the Coriolis force with latitude. The equation becomes:

\[ V_g = \frac{1}{2\Omega \sin \phi} \frac{\Delta p}{d} \]

Solving for \( V_g \), the geostrophic wind, the equation becomes:

\[ V_g = \frac{1}{2\Omega \sin \phi} \frac{\Delta p}{d}. \tag{1} \]

Customarily, the rotational \((2\Omega)\) 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

\[ V_g = \frac{1}{f} \frac{\Delta p}{d}. \]

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/m³. First, we list our data and put them in the proper units, as

\[ \Delta p = 4 \text{ mb} = 400 \text{ Newtons/m}^2 \]
\[ d = 200 \text{ km} = 2 \times 10^6 \text{ m} \]
\[ \sin \phi = \sin(40°) = 0.64 \]
\[ \rho = 0.70 \text{ kg/m}^3 \]
\[ 2\Omega = 14.6 \times 10^{-5} \text{ radian/sec}. \]

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

\[ V_g = \frac{1}{2\Omega \sin \phi} \frac{\Delta p}{d} = \frac{400}{14.6 \times 10^{-5} \times 0.64 \times 0.70 \times 2 \times 10^5} = 30.6 \text{ m/sec}, \text{ or 59.4 knots.} \]

*The rate of the earth's rotation (2\( \Omega \)) 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, 4.18 \times 10^{-5} \text{ degree per second. Most often, } \Omega \text{ is given in radians, where } 2\pi \text{ radians equals } 360° \text{ (} \pi = 3.14). Therefore, the rate of the earth's rotation can be expressed as } 2\pi \text{ radians}=86,154 \text{ sec, or } 7.29 \times 10^{-5} \text{ radian/ sec, and the constant } 2\Omega \text{ becomes } 14.6 \times 10^{-5} \text{ radian/ sec.}
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. 228 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.

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. 9.25a. 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.

Suppose we consider a parcel of air initially at rest at position 1 in Fig. 9.25a. 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 constantly 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

\[
\text{Centripetal force} = \frac{V^2}{r}
\]
FOCUS ON AN OBSERVATION

Estimating Wind Direction and Pressure Patterns Aloft

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 3000 m (about 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 3000 m.

\[
\frac{V^2}{r} + \frac{1}{\rho} \frac{\Delta p}{\rho} + 2\Omega \nu \sin \phi = 0.
\]

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 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. 9.26a) 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. 9.26b, 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. 9.26a. 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.

*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

\[
\frac{1}{\rho} \frac{\Delta p}{\rho},
\]

and the Coriolis force \(2\Omega \nu \sin \phi\). Under these conditions, the gradient wind equation is expressed as

\[
\frac{V^2}{r} + \frac{1}{\rho} \frac{\Delta p}{\rho} + 2\Omega \nu \sin \phi = 0.
\]
In the Southern Hemisphere, the pressure gradient force starts the air moving and the Coriolis force deflects it to the left, thereby causing the wind to blow **clockwise around lows** and **counterclockwise around highs**. Figure 9.27 shows a satellite image of clouds and wind-flow around a low-pressure area in the Northern Hemisphere (9.27a) and in the Southern Hemisphere (9.27b).

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** Figure 9.28 shows a 500-mb chart with contour lines (heavy gray lines), isotherms (dashed red lines), and winds. As we would expect at this level, the regions of lowest elevations (lowest pressures) are associated with the coolest air, and the highest elevations (highest pressures) with the warmest air. As an example, compare the cold −25°C isotherm within the low near the Washington-Oregon coast with the relatively warm −10°C isotherm within the ridge farther to the west.

**FIGURE 9.27**
Clouds and related wind flow patterns around low-pressure areas in (a) the Northern Hemisphere and in (b) the Southern Hemisphere.
The wind directions on the map are given by lines that parallel the wind, whereas the wind speeds are indicated by barbs and flags. Note that the wind generally blows parallel to the contour lines, counterclockwise around the low and its trough and clockwise around the high and the ridge. The strength of the wind is directly related to the magnitude of the contour gradient. (This has the same meaning as the pressure gradient.) Observe the close spacing of the contour lines over the west coast of North America and the corresponding high winds. Near the high-pressure area, where gradients are weak, the winds are light.

In a region where the winds blow in a west-to-east direction, parallel lines of latitude, the wind flow is termed zonal. The winds in Fig. 9.28 are approximately zonal over the eastern third of the United States and Canada. Because the winds aloft generally blow from west to east, planes flying in this direction have a beneficial tail wind, explaining why a flight from San Francisco to New York City takes about 30 minutes less than the return flight. If the flow aloft is zonal, clouds, storms, and surface anticyclones tend to move rapidly from west to east.

When the wind flows in large, looping meanders, following a more north-south trajectory parallel to the meridian lines, the flow is termed meridional. Meridional flow can be seen in Fig. 9.28 along the west coast of North America. Here, cold polar air moves southward on the western side of the low, while relatively warm subtropical air moves northward on its eastern side. As we will see in Chapter 13, during periods of meridional flow, surface storms tend to move slowly, often intensifying into major storm systems.

Moving from south to north, we can see that the contour lines decrease in elevation. We expect this, since the average air temperature is warmer to the south than it is farther north. Where horizontal temperature contrasts are large, there are also large height gradients and strong winds. We can also see that when the flow is zonal, cold air and lower heights are observed to the north of the wind. Where there is meridional flow (such as off the west coast of North America), there is a gradient of height as well as temperature that runs in a west-east direction. In general, the north-south temperature gradient is strongest in winter. This is why the winds aloft are usually much stronger in winter than in summer.

It is clear from Fig. 9.28 that the winds aloft in the middle latitudes of the Northern Hemisphere tend to move in a wavy west-to-east pattern. Because the winds in the Southern Hemisphere blow clockwise around lows and counterclockwise around highs, does this mean that the winds aloft in the middle latitudes of the Southern Hemisphere blow from east to west? The answer is found in the Focus section on p. 231.

Take a minute and look back at Fig. 9.19 on p. 223. 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 (with a pressure gradient of 16 mb per 500 km) 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 geostrophic wind of 55 knots. Why do surface winds normally cross the isobars and why do
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 over the Northern and Southern Hemispheres. Over the equator, where the air is warmer, the pressure aloft is high. 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 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.

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 10.)

In Fig. 9.29a, 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 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 PGF, and the wind blows across the isobars toward lower pressure. The PGF 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. 9.29b). In the Southern Hemisphere, winds blow clockwise and inward around surface lows; counterclockwise and outward around surface highs. See the surface weather map and the general wind flow pattern for South America (Fig. 9.30).

* The geostrophic wind equation given in the Focus section on p. 226 is

\[ V_g = \frac{1}{2 \Omega \sin \phi} \frac{\Delta p}{d} \]

when

- \( p = 16 \text{ mb} = 1600 \text{ Newtons/m}^2 \)
- \( d = 500 \text{ km} = 5 \times 10^8 \text{ m} \)
- \( \sin \phi = \sin(40^\circ) = 0.64 \)
- \( \rho = 1.2 \text{ kg/m}^3 \)
- \( 2 \Omega = 1.4 \times 10^{-5} \text{ radian/sec} \)

The geostrophic wind is

\[ V_g = 1600 \times 16.6 \times 0.64 \times 1.2 \times 5 \times 10^8 = 28.6 \text{ m/sec, or 55 knots.} \]
In Fig. 9.29a, the angle (α) 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, for example, 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.

So far, we have seen that, because of friction, surface winds move more slowly than geostrophic winds 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. As we have seen, if we stand with the wind aloft to our backs, lower pressure will be to our left and higher pressure to our right in the Northern Hemisphere (see Fig. 9.31a). 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. 9.31b and notice that, at the surface, if
we stand with our backs to the wind, then turn clockwise about 30°, lower pressure will be to our left. (In the Southern Hemisphere, if we stand with our backs to the wind, then turn counterclockwise about 30°, lower pressure will be to our right.) 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. As air moves inward toward the center of a low-pressure area (see Fig. 9.32), 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 (that is, move horizontally outward more quickly than it is taken in) to compensate for the converging surface air. As long as the upper-level diverging air balances the converging surface air, the central pressure in the low does not change. However, the surface pressure will change if upper-level divergence and surface convergence are not in balance. For example, if upper-level divergence exceeds surface convergence (that is, more air is removed at the top than is taken in at the surface), the central pressure of the low will decrease, and isobars around the low will become more tightly packed. This 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 (see Fig. 9.32). Again, as long as upper-level converging air balances surface diverging air, the central pressure in 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 we look more closely at the vertical structure of pressure systems in Chapter 13.)

![Diagram](image)

**FIGURE 9.31**
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 (a). At the surface, the same relationship holds if, with your back to the surface wind, you turn clockwise about 30° (b).

**FIGURE 9.32**
Winds and air motions associated with surface highs and lows in the Northern Hemisphere.
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

\[ \frac{1}{\rho} \frac{\Delta p}{\Delta z} + g = 0, \]

where \( \Delta p \) is the decrease in pressure along a small change in height (\( \Delta z \)), and \( g \) is the force of gravity. This expression is usually given as

\[ \Delta p = -\rho g \Delta z. \]

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 \( \rho g \) 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

\[ \Delta p = -\rho g \Delta z, \]

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

Using the hydrostatic equation to compute \( \Delta p \), the difference in pressure in a 1000 m-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 N = 100 Newtons/m²,

\[ \Delta p = 108 \text{ N.} \]

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

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 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 (PGF) 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 PGF 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 (PGF), 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 PGF 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.

The interaction of the forces causes the winds aloft in the Northern Hemisphere to blow clockwise around regions of high pressure and counterclockwise around areas of low pressure. In the Southern Hemisphere, the winds aloft blow 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.

Key Terms

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

- air pressure
- barometer
- millibar
- hectopascal
- mercury barometer
- aneroid barometer
- station pressure
- sea level pressure
- isobars
- constant height chart
- constant pressure (isobaric) chart
- isobaric surface
- contour lines (on isobaric charts)
- ridges
- troughs
- anticyclones
- mid-latitude cyclones
- pressure gradient
- pressure gradient force (PGF)
- Coriolis force
- geostrophic wind
- gradient wind
- centripetal acceleration
- centripetal force
- zonal flow
- meridional flow
- friction layer
- Buys-Ballot’s Law
- hydrostatic equilibrium
- hydrostatic equation

Questions for Review

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