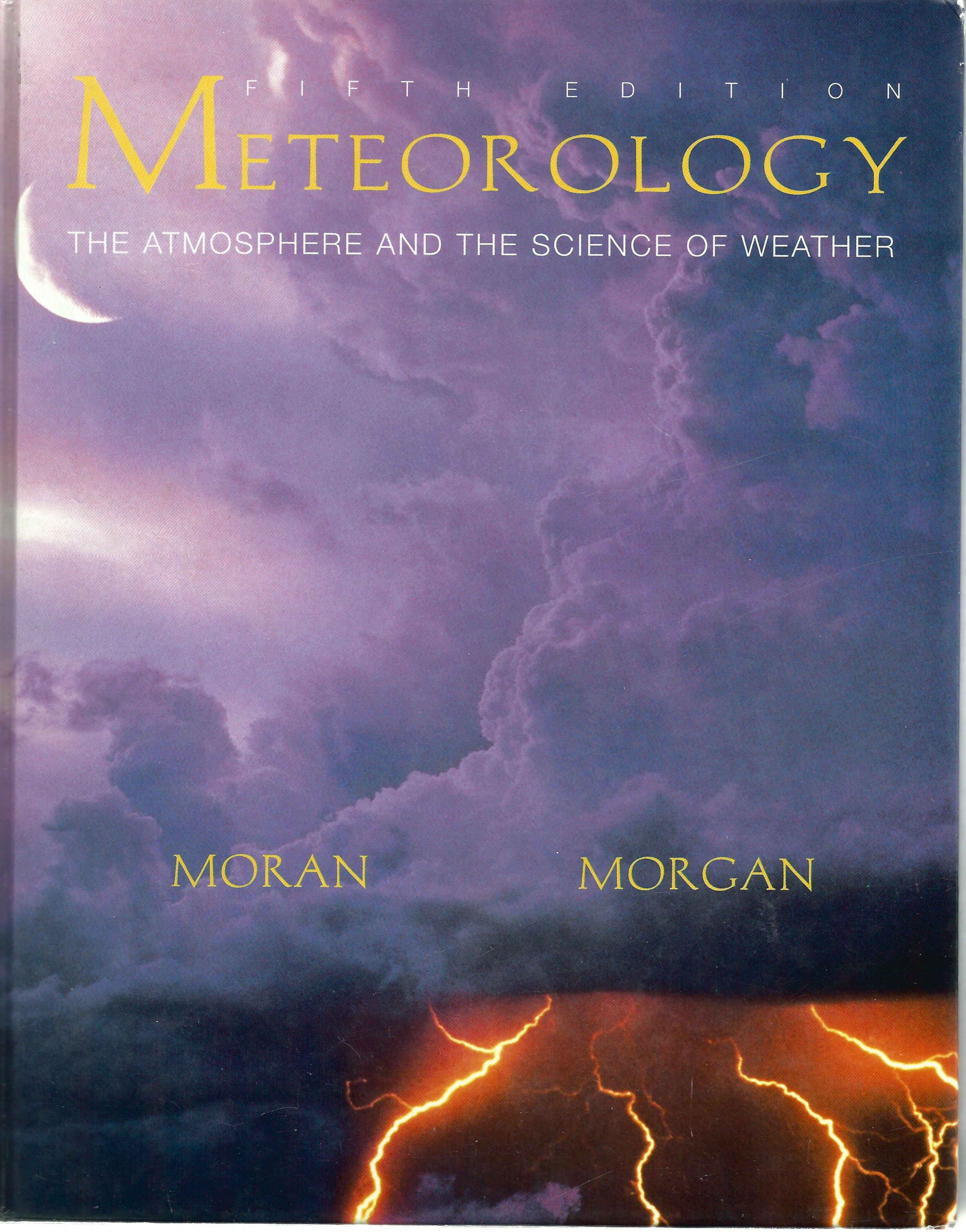

FIFTH EDITION

METEOROLOGY

THE ATMOSPHERE AND THE SCIENCE OF WEATHER

MORAN MORGAN

FIFTH EDITION

Meteorology

The Atmosphere and the Science of Weather

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CHAPTER

9

The Wind

Many forces interact to initiate and shape the wind. The sun is the ultimate source of energy that sustains atmospheric circulation. [Photograph by J. M. Moran]

No one can tell me,
Nobody knows,
Where the wind comes from,
Where the wind goes.

A. A. MILNE,
"Wind on the Hill" Now We Are Six

Some weather systems favor clear skies, light winds, and frosty mornings, whereas others bring ominous clouds, precipitation, and biting winds. Some weather systems trigger brief showers, and others are accompanied by persistent fog and drizzle. Certain weather systems dominate the weather over thousands of square kilometers for prolonged periods. Different weather systems bring different types of weather depending on the air circulation that characterizes each system.

Wind is the motion of air measured relative to the rotating Earth. The atmosphere is coupled to the planet and rotates with it. Once every 24 hr, every point on the Earth's surface and in the atmosphere describes a circular path in space. The circumference of that path decreases with increasing latitude. Hence, at the equator, the atmosphere moves eastward at 1670 km (1035 mi) per hour; at 60 degrees N or S, the speed is 835 km (517 mi) per hour. Why is it that we are not aware of this rapid motion of the air? The reason is that we also move with the rotating planet and its atmosphere at the same speed.

Wind has both magnitude (speed) and direction; that is, wind is a *vector* quantity. A distinction is usually made between the horizontal (east–west and north–south) and the vertical (up–down) components of the wind. Except in small, intense weather systems such as thunderstorms, the magnitude of vertical air motion is typically only 1% to 10% of the horizontal wind speed. But, as we saw in Chapter 6, the vertical component plays the key role in cloud formation. Also, as we will see later in this chapter, vertical and horizontal components of the wind are linked so that a change in one may be accompanied by a change in the other.

In this chapter, we describe the various forces that initiate and control the wind. We begin by examining each force separately as if each force acted independently of all the other forces. We then show how these forces combine to initiate and modify atmospheric circulation. In the Special Topic "Wind Power," we also consider how modern technology is tapping the energy of air in motion.

The Forces

Several forces influence the wind. A *force* is defined as some agent that causes an object at rest to move or that alters the movement of an object that is already in motion. Like the wind, a force has both direction and magnitude (a vector quantity).

For convenience of study, imagine the wind as a continuous stream of air composed of discrete *air parcels*. Assume that any force acting on an air parcel represents the influence of that same force on a stream of air parcels, in other words, on the wind. Now assume also that each parcel consists of a unit mass of air—a single kilogram, for example. Hence, in examining each force that influences air motion, we examine the force per unit mass of air.

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to the [illegible]

A force per unit mass is numerically equivalent to an acceleration. This equivalency follows from **Newton's second law of motion,*** where

force = mass × acceleration

For this reason, we sometimes use the terms *force* and *acceleration* interchangeably when we consider the motion of air parcels. Although the terms are numerically equivalent, a change in velocity is actually a response to a force. A force acts on an air parcel to bring about an acceleration or deceleration of that parcel.

Forces acting on air parcels, which either initiate or modify motion, are the consequence of (1) air pressure gradients, (2) the centripetal force, (3) the Coriolis effect, (4) friction, and (5) gravity. Actually, the centripetal force is not an independent force but a consequence of other forces. Nonetheless, it is instructive to examine the centripetal force along with the other forces.

PRESSURE GRADIENT FORCE

A *gradient* is simply a change in some property with distance. An **air pressure gradient** exists whenever air pressure varies from one place to another. As noted in Chapter 5, spatial variations in air pressure can arise from contrasts in air temperature (principally), from differences in water vapor concentration, or from both. An air pressure gradient thus develops between a mass of cold, dry air and a mass of warm, humid air. In addition, diverging and converging winds can bring about air pressure changes and thereby induce an air pressure gradient. (This important process is covered in Chapter 11, where we examine the origin of highs and lows.)

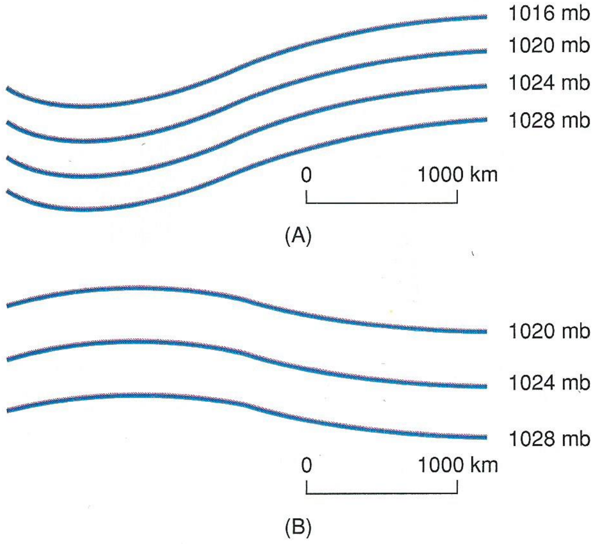
Air pressure gradients develop both horizontally and vertically within the atmosphere. A horizontal pressure gradient refers to air pressure changes along a surface of constant altitude (mean sea level, for example). A vertical pressure gradient is a permanent feature of the atmosphere because air pressure is greatest at the Earth's surface and decreases with altitude.

In order to represent horizontal air pressure gradients on a weather map, the air pressure measured at each weather station is first reduced to sea level, as discussed in Chapter 5. Lines are then drawn joining localities that have the same air pressure reading; some interpolation between weather stations is always necessary. These lines are called **isobars,** and the interval between successive isobars is usually 4 mb.

An isobaric analysis is used to locate centers of high and low pressure and to determine the magnitude of the horizontal air pressure gradient between weather systems. Closely spaced isobars (Figure 9.1A) mean that air pressure changes rapidly with distance, and the pressure gradient is described as steep or strong. More widely spaced isobars (Figure 9.1B) indicate that air pressure changes less with distance, and the pressure gradient is weaker. Note that air pressure gradients are always measured in a direction perpendicular to the isobars.

How do air pressure gradients influence the movement of air? Let us examine an analogous situation. Suppose a bathtub is partially filled with water, as shown in Figure 9.2. As we slosh the water back and forth from one end of the tub to the other, a water pressure gradient develops along the bottom of the tub. At any instant, the water pressure is high where the water level is high, and low where the water level is low. If we stop agitating the water, the water level gradually returns to a horizontal surface and thus creates a uniform water pressure everywhere along the tub bottom. Hence,

FIGURE 9.1
The horizontal air pressure gradient is relatively steep where isobars are close together (A) and weaker where isobars are farther apart (B). Isobars are lines of equal air pressure. Here the contour interval (the difference between successive isobars) is 4 mb.



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*This is the second of three fundamental laws of motion first formulated in 1687 by Sir Isaac Newton.

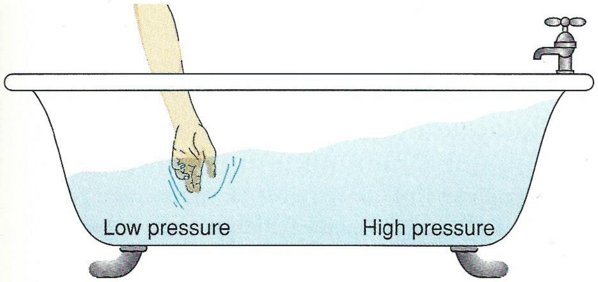


FIGURE 9.2
Sloshing water in a bathtub back and forth creates a horizontal pressure gradient along the bottom of the tub. The water pressure gradient in the tub is analogous to horizontal air pressure gradients in the atmosphere. That is, in response to a pressure gradient, the water (or air) flows from an area of higher pressure toward an area of lower pressure.

in response to a water pressure gradient, water flows from one end of the tub (where the water pressure is greater) to the other end (where the water pressure is less), thus eliminating the pressure gradient.

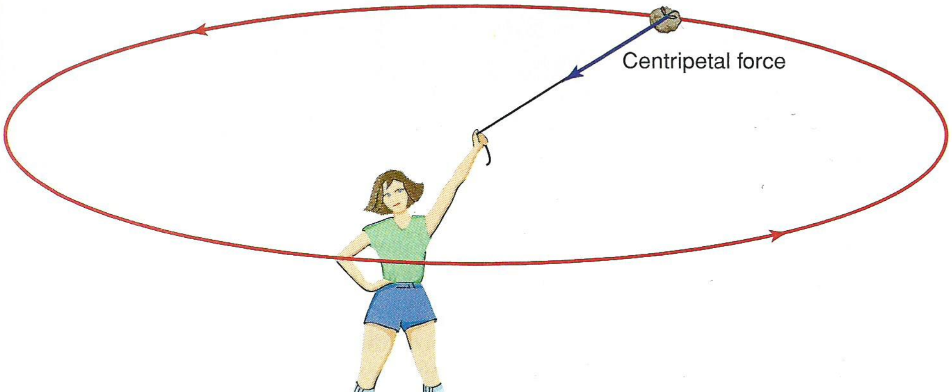
Similarly, when an air pressure gradient develops, air flows in such a way as to eliminate the pressure gradient. Thus, the wind blows away from regions where air pressure is relatively high and toward locales where air pressure is relatively low. The wind is strong where the pressure gradient is steep (closely spaced isobars), and light or calm where the pressure gradient is weak (widely spaced isobars). The force that causes air parcels to move as the consequence of an air pressure gradient is known as the **pressure gradient force.**

CENTRIPETAL FORCE

We may illustrate centripetal force by a simple demonstration. We tie a rock to a string and whirl it about so that the tethered rock describes a circular orbit of constant radius (Figure 9.3). We then cut the string and the rock flies off in a straight line. The behavior of the rock illustrates **Newton's first law of motion;** that is, an object in straight-line, unaccelerated motion remains that way unless acted upon by an unbalanced force. Prior to being cut, the string exerts a net force on the rock by confining it to a curved (circular) path. The net force is directed inward, perpendicular to the direction of motion, and toward the center of the circular orbit. For this reason, the net force is known as the **centripetal** *(center-seeking)* **force.** When we cut the string, the centripetal force no longer operates and the rock follows a straight path.

A net force causes an acceleration. We usually think of an acceleration as a change in speed, as when an automobile speeds up or slows down. But acceleration is a vector quantity; that is, it has both magnitude and direction. Hence, an acceleration may consist of a change in either speed or direction, or both. In our rock-on-a-string example, centripetal force is responsible only for a continuous change in the direction of the rock

FIGURE 9.3
A rock attached to a string describes a circular path. Centripetal (for "center-seeking") is the name we give to the force that confines an object to a curved path. If the string is cut, the centripetal force is eliminated and the rock flies off in a straight line (tangent to the circular path).



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SPECIAL TOPIC

Wind Power

Harnessing the energy of the wind is a technology that was well established as early as the twelfth century in portions of the Middle East where water power was not available. In North America, the energy crisis of the 1970s spurred renewed interest in this ancient technology. Today scientists are employing modern aerodynamic principles and space-age materials in designing and constructing modern wind-driven turbines that convert some of the wind's kinetic energy into electricity (Figure 1).

In Chapter 4, we saw how the sun drives the atmosphere. Although only about 2% of the solar energy that reaches the Earth is ultimately converted to the kinetic energy of wind, that is still a tremendous quantity of energy. Theoretically, windmill blades can convert a maximum of 60% of the wind's energy into mechanical energy. In practice, however, wind generators extract only about 25% of the wind's energy. Furthermore, average wind speeds must be at least 19 km (12 mi) per hour before most wind-powered electricity-generating systems can operate economically.

The power that a windmill can extract from the wind is directly proportional to (1) air density, (2) the area swept out by the windmill blade, and (3) the cube of the instantaneous wind speed (v^3). Wind speed is by far the most important consideration in evaluating a region's wind energy potential. Even small changes in wind speed translate into large changes in energy output. For example, doubling the wind speed (a common occurrence) multiplies the available wind power by a factor of eight (2 × 2 × 2).

Tapping the wind's energy faces several difficulties. Both wind speed and direction vary continuously with time (see Figure 9.27), and wind speed also varies with the exposure of the site, roughness of the terrain, elevation above the surface, and season of the year. As a general rule, a minimum of several years of detailed wind monitoring is needed for a preliminary evaluation of wind power potential at any site. The long-term

FIGURE 1
A wind turbine that feeds electricity into the local grid. [Photograph by J. M. Moran]

(a curved rather than straight-line path); the rock neither speeds up nor slows down.*

The centripetal force is not itself an independent force; rather, it arises from the action of other forces and may be the consequence of imbalances in other forces. In our rock-on-a-string example, the string is responsible for the centripetal force. Consider another example. Suppose that you are a passenger in an auto that rounds a curve at a high rate of speed. You feel a force that pushes you outward from the curve. Actually, what you experience is the tendency of your body to continue in a straight path while the auto is

---

*It can be demonstrated that the acceleration imparted to a unit mass by the centripetal force is directly proportional to the square of the velocity and inversely proportional to the radius of curvature.

climate record should also be consulted to check for the frequency of potentially destructive winds. Wind data from nearby weather stations can be very useful, as long as care is taken in extrapolating wind data between localities and from one elevation to another.

At ordinary speeds, wind is a relatively diffuse energy source, comparable in magnitude to insolation. Hence, a wind turbine's power generation potential also depends on the area swept out by the windmill blades. Larger windmill blades harvest more energy. Until recently, design and strength-of-materials problems limited the size of wind turbines. Today, use of stronger and lighter materials for blades are making possible larger wind turbines that generate as much as 0.5 megawatt of electricity.

The most formidable obstacle to the development of wind power potential stems from the inherent variability of the wind. The electrical output of wind turbines varies as a consequence, and a wind power system must include a means of storing the energy generated during gusty periods for use when the wind is light or calm. A 3- to 5-kW wind turbine is needed to meet the total electrical requirements, including heating, of a typical household. Today, such systems are commercially available, but the cost of materials and construction—including a tower, storage batteries, and generator—ranges from about $5,000 to $20,000.

Economy of scale dictates that centralized arrays of many wind turbines, called *windmill farms,* are preferable to individual household wind turbines. Windmill farms consist of 50 or more super wind-power generators, each capable of producing as much as 0.5 megawatt of electricity. Windmill arrays currently operating in California supplement conventional energy sources by feeding electricity directly to existing power grids. Since 1981, more than 16,000 turbines have been installed in windy mountain passes in California. By 1995, California's windmill farms were supplying about 1.5% of the state's total electrical needs; this figure is expected to climb to 10% by 2005.

In view of current technological and economic limitations, wind power has its greatest immediate potential in regions where average winds are relatively strong and consistent in direction. In North America, such regions include the western High Plains, the Pacific Northwest coast, portions of coastal California, the eastern Great Lakes, the south coast of Texas, and exposed summits and passes in the Rockies and Appalachians. In fact, by one estimate, the wind power potential of North and South Dakota could meet 80% of total U. S. electrical needs.

Recent technological advances and tax incentives are beginning to make wind power more cost-competitive with conventional methods of energy generation. Between 1980 and 1993, the installed cost of new wind turbines dropped from about $4000 per kilowatt to about $1000 per kilowatt. U. S. Windpower, the nation's major wind turbine manufacturer, has developed a turbine that operates over a wind-speed range of 10 to 97 km (6 to 60 mi) per hour. Earlier models were limited to a range of about 19 to 73 km (12 to 45 mi) per hour. Also, the 1992 Federal Energy Bill granted wind turbine operators a special tax break of $0.015 per kilowatt-hour. By the close of the century, many electric utilities are expected to increase their exploitation of wind power potential.

following a curved path. In this case, the frictional resistance of the tires against the pavement provides the centripetal force.

It follows from this discussion that a centripetal force operates whenever the wind follows a curved path. As we will see later in this chapter, however, the centripetal force results from an imbalance of other forces operating in the atmosphere.

CORIOLIS EFFECT

Imagine that you are located far away at some fixed point in space and are looking back at planet Earth. Over many hours, you follow the track of a storm, which is clearly identifiable by its slowly swirling mass of clouds (Figure 9.4). From your perspective, the storm center appears to move in a straight line at constant

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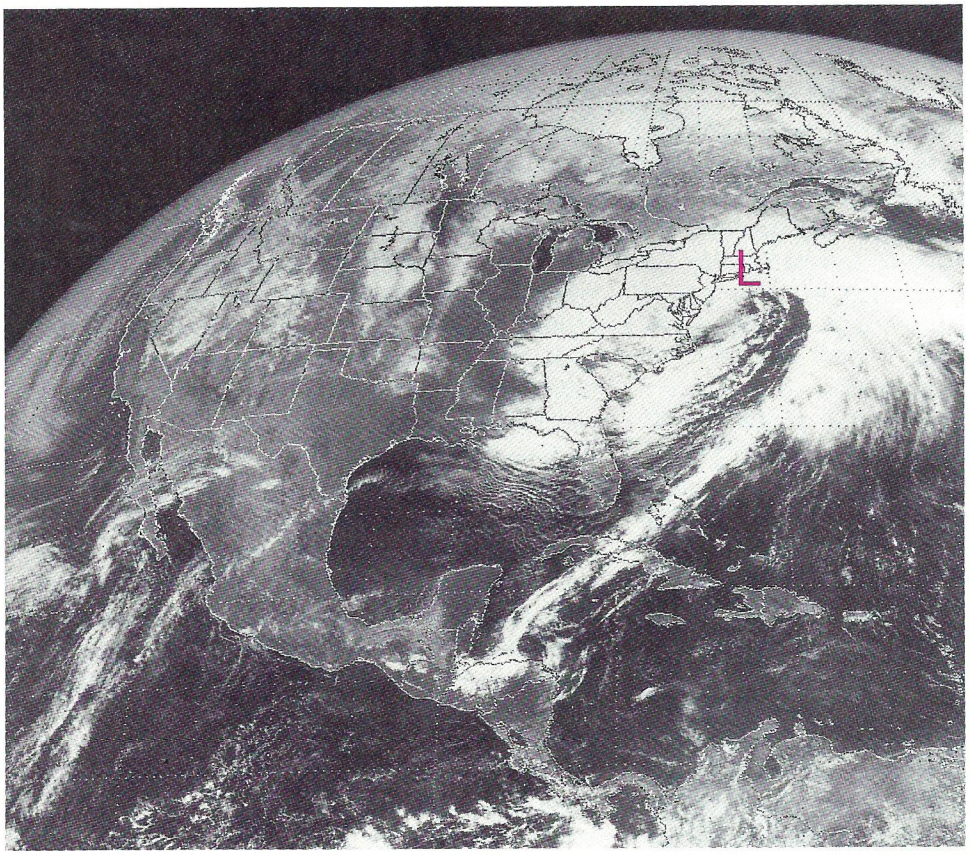

FIGURE 9.4
In this visible satellite image, an intense midlatitude cyclone is readily identified by its swirling mass of clouds. The storm center is located just east of New Jersey. [NOAA National Environmental Satellite, Data, and Information Service]

speed. At the same time, an observer on Earth is tracking the storm. From that observer's perspective, the storm center appears to follow a curved path. Surprisingly, both descriptions of the storm's track are correct!

The two descriptions are correct because the two observers used different frames of reference in following the storm's movements. The Earthbound observer's frame of reference is the familiar north–south, east–west, and up–down coordinate system that rotates with the Earth. To the Earthbound observer, it is not obvious that this coordinate system is rotating because it and the observer rotate together. Viewed from space, however, the Earthbound coordinate system actually shifts as the Earth rotates (Figure 9.5). It is as if the Earth and the coordinate system rotate under the storm (or any other object moving over the Earth's surface). Meanwhile, from your distant vantage point in space, you followed the storm's movement with respect to a nonrotating coordinate system, fixed in space. In summary, then, the difference in storm track (straight versus curved) is due to the difference in coordinate system (nonrotating versus rotating).

Recall our earlier discussion of Newton's first law of motion and the centripetal force. We saw that curved motion implies that a net force is operating, whereas unaccelerated, straight-line motion implies a balance of forces. If we apply this law to our storm track example, we conclude that a net force operates when we use the Earthbound rotating coordinate system, whereas forces are balanced when we use the nonrotating coordinate system fixed in space. Hence, changing our frame of reference (coordinate system) from nonrotating to rotating gives rise to a net force responsible for curved motion. This deflective force is known as the **Coriolis effect,** named for Gaspard Gustav de Coriolis, who first described the phenomenon mathematically in 1835.

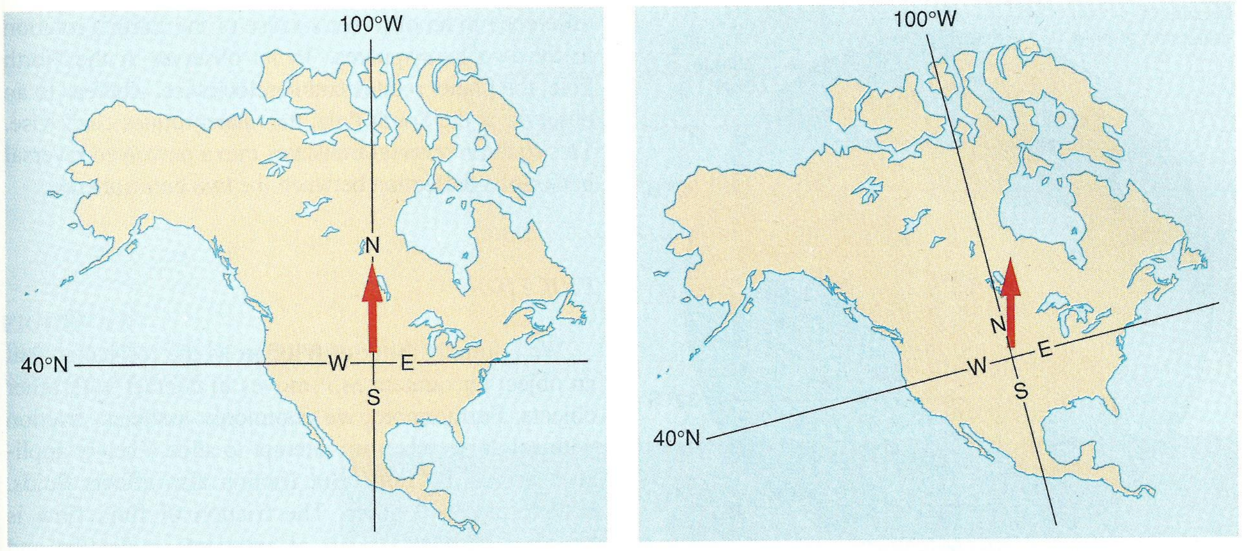

FIGURE 9.5
When viewed from space, our north–south, east–west frame of reference changes as Earth rotates on its axis. Here a south wind in the Northern Hemisphere is deflected (relative to the frame of reference) to the right and becomes a southwest wind. [Adapted from F. K. Lutgens and E. J. Tarbuck, *The Atmosphere: An Introduction to Meteorology, 4th ed.,* © 1989, p. 169. Adapted by permission of Prentice-Hall, Inc., Upper Saddle River, NJ]

Wind direction and wind speed are measured with respect to the north–south, east–west, and up–down frame of reference that rotates with the planet. Therefore, we must take the Coriolis effect into account in any explanation of air circulation. Because of the Coriolis effect, the wind is deflected to the right of its initial direction in the Northern Hemisphere and to the left of its initial direction in the Southern Hemisphere (Figure 9.6).

Although the Coriolis effect influences the wind regardless of its direction, the magnitude of the effect varies significantly with latitude.* This is because the Coriolis effect stems from the rotation of the Earth on its axis, which imparts a rotation to our Earthbound frame of reference. The rotation of our frame of reference is maximum at the poles and declines with decreasing latitude to zero at the equator. This variation can be understood by visualizing the daily rotation of towers situated at different latitudes. In a 24-hr day, the Earth makes one complete rotation, as would a tower situated at the North or South Pole. In the same period, a tower at the equator would not rotate at all, but instead would describe an end-over-end motion. For a tower located at any latitude in between, some rotation of the tower occurs as the Earth rotates but not as much as at the poles. The Coriolis effect is thus latitude-dependent: the Coriolis effect is zero at the equator and increases with latitude to a maximum at the poles.

The Coriolis effect also varies with wind speed; that is, the deflection increases as the wind strengthens. This is because, in the same period of time, fast air parcels cover greater distances than slow air parcels. The longer the trajectory, the greater is the shift of the coordinate system with respect to a moving air parcel. For practical purposes, the Coriolis effect has an important influence only on the air circulation within large-scale weather systems, that is, systems larger than thunderstorms.

A rotational motion usually accompanies the draining of water from a sink or a bathtub. A popular misconception is that the direction of this rotation (clockwise or counterclockwise) is consistently in one direction in the Northern Hemisphere and in the opposite direction in the Southern Hemisphere, presumably because of the Coriolis effect. At the very small scale represented by the sink or bathtub, however, the magnitude of the Coriolis effect is simply too small to have a significant influence on the direction of rotation. The drainage direction is more likely a consequence of some residual motion of the water when the sink or bathtub was first

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*The magnitude of the Coriolis effect varies with the sine of the latitude, increasing from 0 at the equator (0 degrees latitude) to a maximum at 90 degrees (the North and South poles).

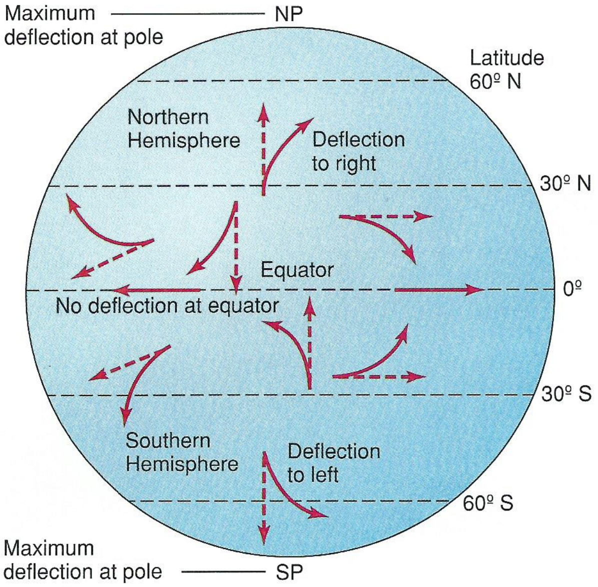

FIGURE 9.6
Large-scale winds are deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. This deflection is due to the Coriolis effect.

filled with water and can be either clockwise or counterclockwise.

It is useful to know why the Coriolis effect reverses direction between the hemispheres and causes large-scale winds in the Southern Hemisphere to swerve to the left rather than to the right. This reversal is related to the difference in an observer's sense of the Earth's rotation in the two hemispheres. To an observer at the North Pole, the planet rotates counterclockwise, whereas to an observer at the South Pole, the planet rotates clockwise. This rotation reversal translates into a perceived reversal in Coriolis deflection between the two hemispheres.

FRICTION

We normally think of **friction** as the resistance that an object encounters as it moves in contact with other objects. Furthermore, we commonly associate friction with solids, as when we attempt to slide a heavy appliance across the floor. But friction also affects fluids, both liquids and gases. The friction of fluid flow is known as **viscosity** and is of two types, *molecular* and *eddy.*

One source of fluid friction is the random motion of molecules composing a liquid or gas; this type of fluid friction is called **molecular viscosity.** Considerably more important, however, is fluid friction that arises from much larger irregular motions, called eddies, which develop within fluids; this type of fluid friction is known as **eddy viscosity.**

The stream in Figure 9.7 illustrates the effects of eddy viscosity. Rocks in the streambed obstruct the flow of water and cause the current to break into eddies to the lee of the rocks. Eddies, visible as white water, tap some

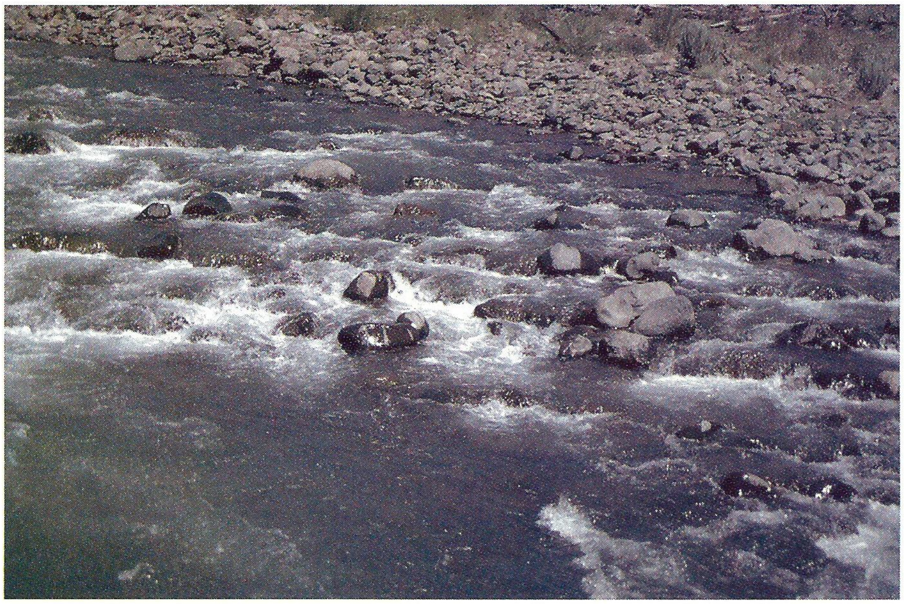

FIGURE 9.7
Rocks in a riverbed break the current into turbulent eddies that appear as white water to the lee of rocks. [Photograph by J. M. Moran]

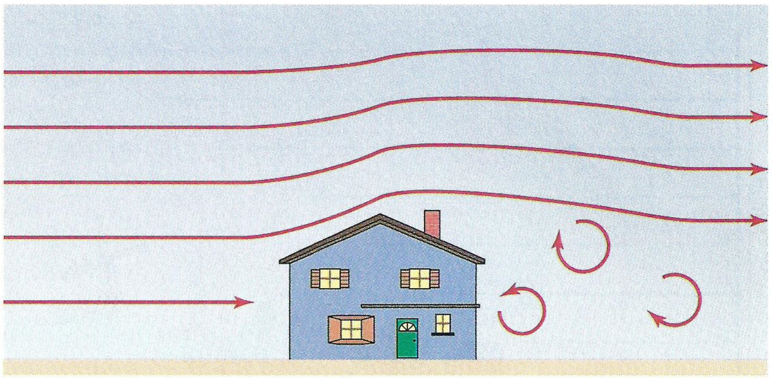

FIGURE 9.8
Turbulent eddies develop in the wind on the leeward side of a house.

FIGURE 9.9
A snow fence slows the wind, thereby decreasing the wind's ability to transport snow. Hence, snow accumulates to the lee of the snow fence. [Photograph by J. M. Moran]

of the stream's kinetic energy so that the stream slows. In an analogous manner, obstacles on the Earth's surface such as trees, houses, and telephone poles break the wind into eddies of various sizes to the lee of each obstacle (Figure 9.8). Consequently, the wind slows.

A snow fence (Figure 9.9) provides a practical illustration of the frictional slowing of the wind. Snow fences are designed to trap wind-blown snow, in some instances to prevent snow from drifting onto a nearby highway and in others to keep the soil snow covered.* A snow fence breaks the wind into small eddies, thereby tapping some of the wind's kinetic energy. The wind diminishes and loses some of its snow-transporting ability. Consequently, snow accumulates on the downwind side of the fence.

The rougher the surface of the Earth, the greater is the eddy viscosity of the wind. A forest thus offers more frictional resistance to the wind than does the smoother surface of a freshly mowed lawn. The eddy viscosity diminishes rapidly with altitude above the Earth's surface, away from the obstacles mainly responsible for frictional resistance. Hence, wind speed increases with altitude. This explains the advantage of siting a windmill at as high an elevation as possible (Figure 9.10). Above an average altitude of about 1 km (0.62 mi), friction is a minor force that merely acts to smooth the flow of air. The atmospheric zone to which frictional resistance (eddy viscosity) is essentially confined is called the **friction layer.**

Turbulence is fluid flow that is characterized by eddy motion. So far we have been considering obstacles on the Earth's surface as sources of eddies that exert a drag on the wind. Irregular fluid flow that originates in this way is known as *mechanical turbulence*. In addition,

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*Snow accumulated downwind of a snow fence protects the underlying soil from deep penetration of subfreezing temperatures. The snow accumulation also ensures a supply of soil moisture.

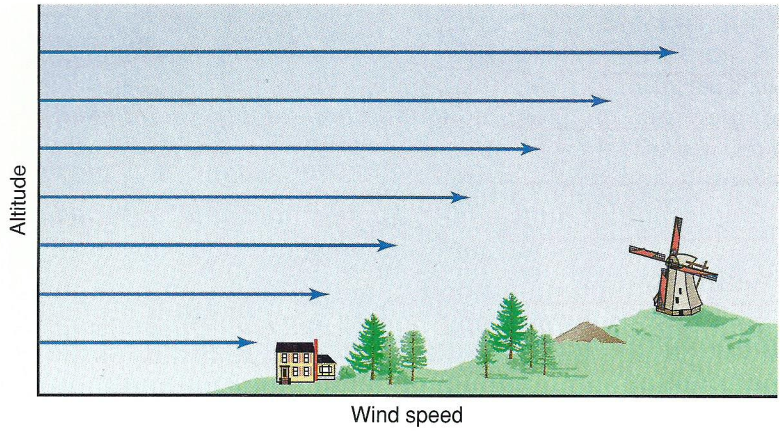

FIGURE 9.10
The horizontal wind strengthens with altitude, away from the frictional resistance offered by objects on the Earth's surface.

eddies develop in air as a consequence of solar heating of the ground. Irregular fluid flow that originates in this way is known as *thermal turbulence.* Convection currents are examples of thermal turbulence within the atmosphere. In actual practice, it is virtually impossible to distinguish between the two sources of turbulent eddies.

Regardless of source, we experience turbulent eddies as gusts of wind. You may have noticed that the gustiness of the wind often varies with the time of day; that is, gusts tend to be strongest during the warmest hours of the day. The reasons for this relationship are discussed in the Special Topic "Wind Gusts, Wind Shear, and Atmospheric Stability" later in this chapter.

GRAVITY

The atmosphere is subject to the same force that holds all objects on the Earth's surface, **gravity.** The force of gravity is actually the net effect of two other forces working together: (1) the force of attraction between the Earth and all other objects, called **gravitation,** and (2) a much weaker centripetal force imparted to all objects because of their rotation with the Earth on its axis. The two forces combine to produce the force of gravity, which accelerates a unit mass of any object downward at the rate of 9.8 m per sec each second.

The force of gravity always acts downward and perpendicular to the Earth's horizontal surface. For this reason, gravity, unlike the Coriolis effect and frictional forces, does not modify the horizontal wind. Gravity does, however, affect air that is ascending or descending, such as in convection currents, and gravity is responsible for the downhill drainage of cold, dense air.

SUMMARY

We have now examined the various forces that affect horizontal and vertical air motion, and we can draw the following conclusions.

1. The *pressure gradient force* accelerates air away from regions of high air pressure and toward areas of low air pressure. Acceleration is directly proportional to the pressure gradient; that is, the closer the spacing of isobars, the greater is the acceleration.
2. A *centripetal force* is an imbalance of actual forces and operates whenever the wind describes a curved path. It is responsible for a change in wind direction, but not wind speed.
3. The *Coriolis effect* deflects large-scale winds to the right of the initial direction in the Northern Hemisphere and to the left of the initial direction in the Southern Hemisphere. Its magnitude varies from zero at the equator to a maximum at the poles and is directly proportional to wind speed.
4. *Friction* acts opposite the wind direction and slows winds within about 1 km (0.62 mi) of the Earth's surface.
5. *Gravity* accelerates air downward toward the Earth's surface but it does not modify horizontal winds.

Joining Forces

To this point, we have examined forces operating in the atmosphere as if each force acted independently of all the others. In reality, these forces interact with one another in governing both the direction and speed of the wind. In some cases, two or more forces achieve a balance or equilibrium. From Newton's first law of motion, when the forces acting on an air parcel are in balance, there is no net force, and the parcel either remains stationary or continues to move along a straight path at constant speed. Therefore, when forces balance, the net acceleration is zero.

Let us now examine how forces interact in the atmosphere to control the vertical and horizontal flow of air, that is, the wind. These interactions result in (1) hydrostatic equilibrium, (2) the geostrophic wind, (3) the gradient wind, and (4) surface winds, the winds within the friction layer (Table 9.1).

HYDROSTATIC EQUILIBRIUM

We noted earlier that the atmosphere features a vertical pressure gradient. As shown schematically in Figure 9.11, the force due to this pressure gradient is directed upward from high pressure at the Earth's surface toward lower air pressure aloft. If this force acted alone, the vertical pressure gradient force would accelerate air away from the Earth, and we would be left gasping for breath. However, except in some small-scale violent weather systems (tornadoes, for example), the atmosphere's vertical pressure gradient force is almost balanced by the equal but oppositely directed force of gravity. An actual balance of these two forces is known as **hydrostatic equilibrium.**

Whenever forces are in balance, no net acceleration occurs; that is, there is no change in velocity. Hydrostatic equilibrium, then, does not preclude vertical (up or down) motion of air. Because of balanced forces, upward-moving air parcels continue upward at *constant* velocity, and downward-moving air parcels continue downward at *constant* velocity. Slight deviations from hydrostatic equilibrium, however, cause air to change speed (accelerate) vertically.

TABLE 9.1
Forces Involved in Large-Scale Atmospheric Circulation—A Summary

Forces	*Hydrostatic equilibrium*	*Geostrophic wind*	*Gradient wind*	*Surface winds*
Pressure gradient	X	X	X	X
Centripetal			X	X
Coriolis		X	X	X
Friction				X
Gravity	X			

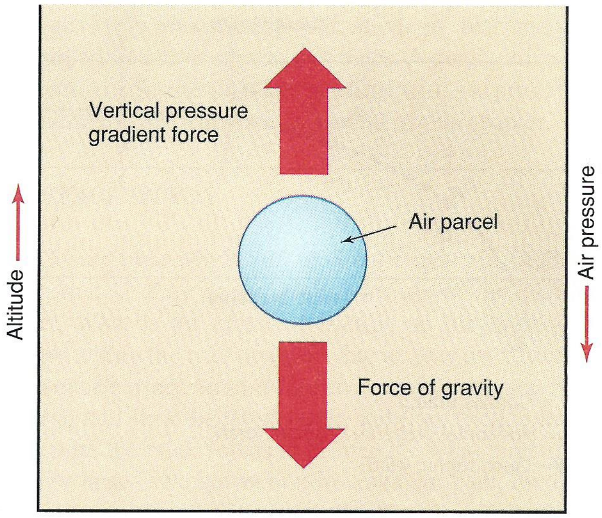

FIGURE 9.11
With hydrostatic equilibrium, the upward-directed vertical pressure gradient force acting on an air parcel is balanced by the downward-directed force of gravity.

GEOSTROPHIC WIND

The **geostrophic wind** is an unaccelerated horizontal wind that flows along a straight path at altitudes above the friction layer. It results from a balance between the horizontal pressure gradient force and the force due to the Coriolis effect.

When a horizontal pressure gradient *(P_H)* develops, air parcels at first accelerate directly across isobars, away from high pressure and toward low pressure (Figure 9.12). As air parcels speed up, however, the Coriolis effect *(C)* strengthens and causes air parcels to swerve gradually to the right of their initial flow direction (in the Northern Hemisphere). The two forces eventually attain a balance, so that the wind blows at a constant speed in a straight path parallel to the isobars with the lowest air pressure to the left of air motion. Because the Coriolis effect is a large-scale phenomenon, the geostrophic wind develops only in large-scale weather systems.

GRADIENT WIND

The **gradient wind** has many characteristics in common with the geostrophic wind. It also is large-scale,

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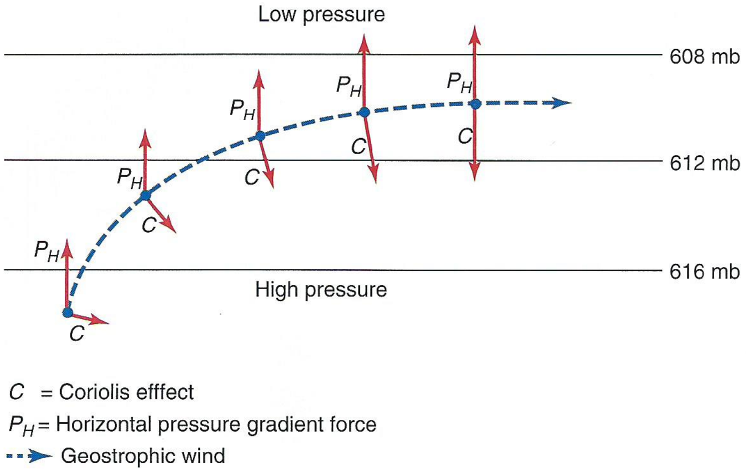


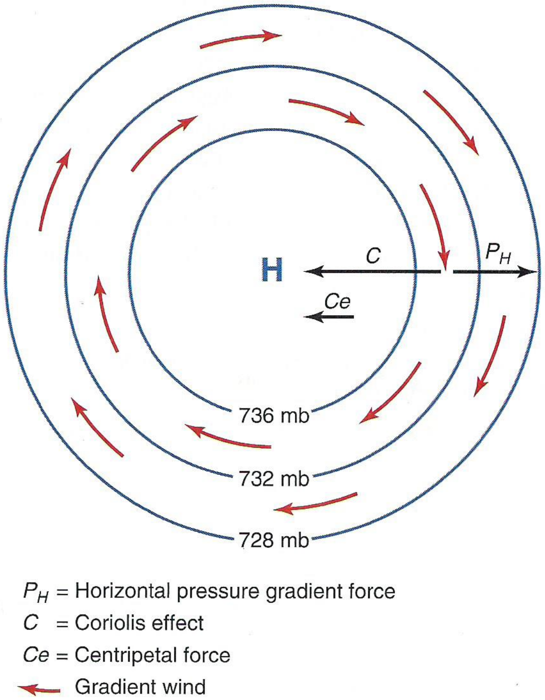
FIGURE 9.12
The horizontal air pressure gradient causes air parcels to accelerate across isobars from areas of high pressure toward areas of low pressure. The Coriolis effect then deflects air parcels to the right in the Northern Hemisphere. The Coriolis effect increases in magnitude until it balances the pressure gradient force. The result is an unaccelerated horizontal wind blowing parallel to isobars, that is, the geostrophic wind.

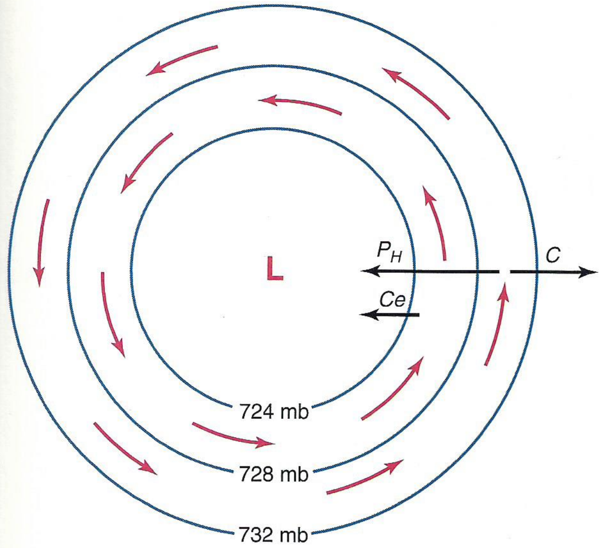
horizontal, and frictionless and blows parallel to isobars. The important distinction between the two is that the geostrophic wind blows in a straight path, whereas the path of the gradient wind is curved. Because a net centripetal force constrains air parcels to a curved trajectory, the gradient wind is not the consequence of balanced forces. Recall that the centripetal force changes only the air parcel's direction of motion, not its speed. The horizontal pressure gradient force, the Coriolis effect, and the centripetal force thus interact in the gradient wind.

A gradient wind develops at altitudes above the friction layer around a dome of high air pressure, called an **anticyclone** (or *high*), or around a center of low air pressure, called a **cyclone** (or *low*). In an ideal anticyclone, isobars form a series of concentric circles about the location of highest air pressure, as shown in Figure 9.13. The horizontal pressure gradient force (P_H) is directed radially outward, away from the center of the high. The Coriolis effect *(C)* is directed inward. The Coriolis effect is slightly greater than the pressure gradient force, with the difference giving rise to the inward-directed centripetal force *(Ce)*. (This is what was meant earlier when we indicated that a centripetal force results from an imbalance of other forces.) In a Northern Hemisphere anticyclone above the friction layer, the gradient wind consequently blows clockwise and parallel to isobars.

In an ideal cyclone, isobars form a series of concentric circles about the location of lowest air pressure. As indicated in Figure 9.14, the horizontal pressure gradient force (P_H) is directed inward toward the cyclone center, and the Coriolis effect *(C)* is directed radially outward from the center of the low. The pressure gradient force is slightly greater than the Coriolis effect, with the difference being equal to the net inward-directed centripetal force *(Ce)*. In a Northern Hemisphere cyclone above the friction layer, the gradient wind consequently blows counterclockwise and parallel to isobars.

FIGURE 9.13
In a Northern Hemisphere anticyclone above the friction layer, the gradient wind blows clockwise and parallel to isobars. In this idealized situation, the isobars form a series of concentric circles.





P_H = Horizontal pressure gradient force
C = Coriolis effect
Ce = Centripetal force
→ Gradient wind

FIGURE 9.14
In a Northern Hemisphere cyclone above the friction layer, the gradient wind blows counterclockwise and parallel to isobars. In this idealized situation, the isobars form a series of concentric circles.

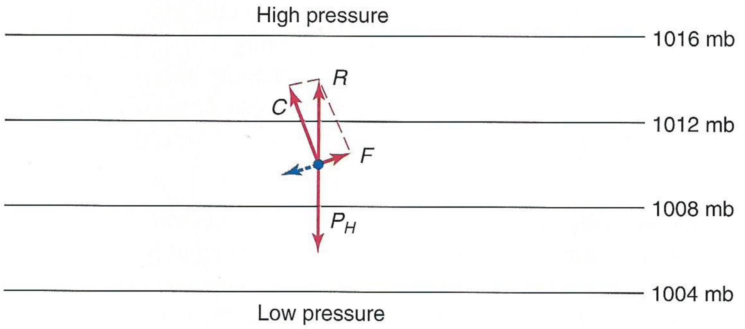
The geostrophic and gradient winds are models (Chapter 1) that only approximate the actual behavior of horizontal winds above the friction layer. These approximations are nonetheless quite useful, and meteorologists rely on such approximations in their analysis of isobaric patterns on weather maps. A quantitative description of geostrophic and gradient winds is presented in the Mathematical Note at the end of this chapter.

SURFACE WINDS

Geostrophic winds and gradient winds are frictionless; that is, they occur at altitudes above the friction layer. What is the effect of friction on the horizontal winds within the friction layer; that is, how does friction influence surface winds? Intuitively, we know that friction should slow the wind, but in addition, friction interacts with the other forces to change the wind direction.

For large-scale air motion in a straight path, the frictional force *(F)* combines with the Coriolis effect *(C)* to balance the horizontal pressure gradient force *(P_H)*, as shown in Figure 9.15. Friction acts directly opposite (180 degrees) the direction of air motion and slows the wind. The Coriolis is at a right angle (90 degrees) to the wind direction and weakens with decreasing wind speed. The Coriolis no longer balances the horizontal pressure gradient force. As shown in Figure 9.15, the horizontal pressure gradient *(P_H)* is balanced by the resultant *(R)* of the Coriolis *(C)* and friction *(F)*. Consequently, friction (due to the roughness of the Earth's surface) causes the horizontal wind to slow down and shift direction across isobars and toward low pressure. The deflection angle of surface winds crossing isobars varies from about 10 degrees over relatively smooth surfaces, where friction is low, to almost 45 degrees over rough terrain, where friction is greater.

FIGURE 9.15
At the Earth's surface and elsewhere within the friction layer, the Coriolis effect combines with friction to balance the horizontal pressure gradient force. As a consequence, the horizontal wind blows across isobars and toward lower air pressure. Recall that the Coriolis effect acts at right angles to the initial wind direction and friction acts in a direction opposite that of the wind.



C = Coriolis efffect
F = Force of friction
R = Coriolis effect + force of friction
P_H = Horizontal pressure gradient force
⇠ Direction of surface wind

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SPECIAL TOPIC

Wind Gusts, Wind Shear, and Atmospheric Stability

You may have noticed that the day's strongest and gustiest winds often occur during the afternoon hours. This phenomenon is perhaps most apparent following a night of extreme radiational cooling. By dawn, a temperature inversion develops in the lowest air layer, the surface air temperature has reached its diurnal minimum, and surface winds are very light or calm. In the hours following sunrise, the surface temperature rises steadily in response to increasing insolation, and through sensible heating, eventually the lowest air layer is destabilized. By late morning or early afternoon, surface winds strengthen and become gusty, only to die down by sunset. These observations suggest a relationship between the gustiness of the wind and atmospheric stability.

Wind gusts are caused by irregular whirls of air called eddies. We are all familiar with the analogous whirls or eddies that form in rapidly flowing streams of water (see Figure 9.7). Eddies characterize turbulent motion of fluids and, as noted elsewhere in this chapter, are either mechanical or thermal in origin.

As a rule, as the wind strengthens, it becomes more turbulent and eddies become more energetic. Within the friction layer, horizontal wind speeds increase with altitude. Hence, strong eddies develop aloft, where winds are strong, and weaker eddies form near the Earth's surface, where winds are lighter. Eddies are transported upward and downward within the friction layer depending on atmospheric stability.

As shown in Chapter 6, atmospheric stability influences the vertical motion of air; that is, stable air suppresses vertical motion, and unstable air enhances vertical motion. When the friction layer is unstable, strong eddies are transported from higher altitudes downward toward the surface. Consequently, winds aloft weaken because they lose some kinetic energy, and winds at the surface strengthen and become gusty because they gain some kinetic energy from eddies descending from aloft. On the other hand, when the friction layer is stable, strong eddies generated aloft remain there, and weak eddies that form near the surface remain near the surface.

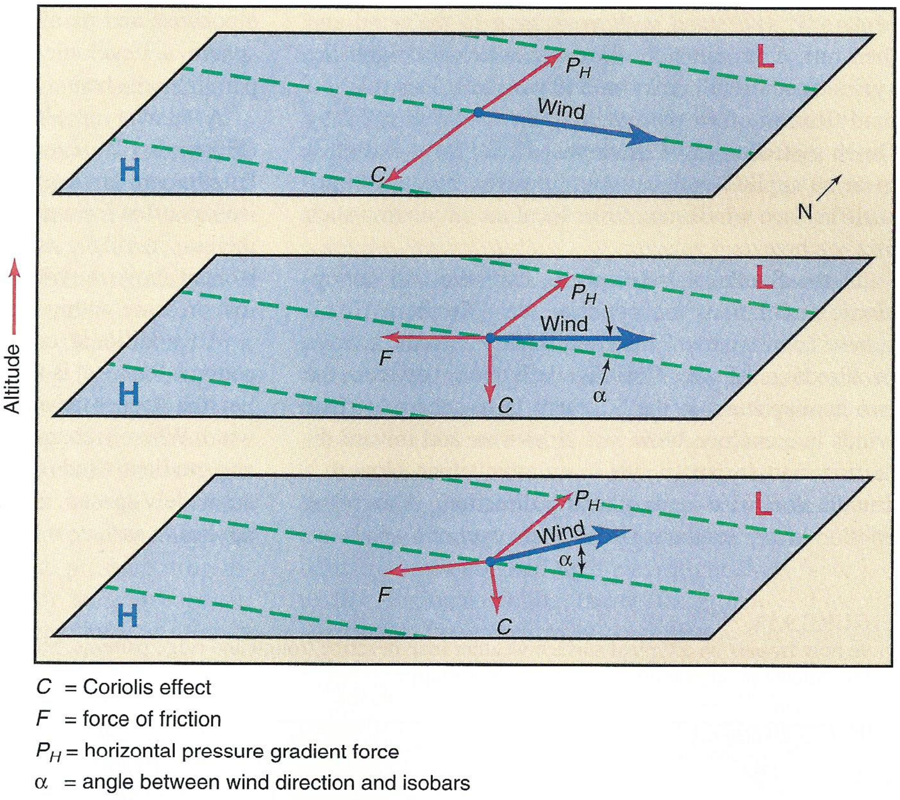
It follows that vertical wind shear and stability are related. A *wind shear* occurs whenever the wind changes speed (or direction) with distance or, as in this case, with altitude. Hence, a vertical wind shear characterizes the friction layer because horizontal winds strengthen with altitude. The magnitude of this wind shear, however, varies and is influenced by air stability. As noted, when the air is unstable, eddy transport reduces the vertical wind shear (the difference in horizontal wind speed between the surface and aloft). When the air is stable, the lack of eddy transport increases the vertical wind shear. In fact, although winds are typically light or calm within a low-level temperature inversion, winds are often quite strong in the air layer just above the inversion.

In summary, the greater the stability of the atmosphere's friction layer, the stronger is the vertical wind shear and the less gusty are surface winds. An unstable friction layer features a weaker vertical wind shear and relatively energetic and gusty surface winds. Thus, in our example, as the lower troposphere is destabilized in the hours following sunrise, surface winds strengthen and become gusty.

As we noted earlier, friction's influence on the horizontal wind diminishes with altitude and is negligibly small at the top of the friction layer. Hence, horizontal winds strengthen with altitude. Furthermore, the angle between the wind direction and the isobars is maximum near the Earth's surface, decreases with altitude, and is essentially zero at the top of the friction layer (Figure 9.16).

What is the effect of friction on the horizontal surface winds blowing in an anticyclone and cyclone? As with straight-line surface winds, friction slows cyclonic and anticyclonic winds and combines with the Coriolis effect to shift winds so that they blow across isobars and toward low pressure. At the Earth's surface, therefore, anticyclonic winds blow clockwise and outward, as shown in Figure 9.17, and surface cyclonic winds blow counterclockwise and inward, as shown in Figure 9.18.

The characteristic circulation of a cyclone enables us to formulate a simple rule of thumb for locating cyclone



C = Coriolis effect
F = force of friction
P_H = horizontal pressure gradient force
α = angle between wind direction and isobars

FIGURE 9.16
For the same horizontal pressure gradient, the angle between the wind direction and the isobars (dashed lines) decreases with altitude. [Adapted from R. A. Anthes, *Meteorology, 6th ed.,* New York: Macmillan Publishing Company, 1992, p. 79]

FIGURE 9.17
Viewed from above, surface winds blow clockwise and outward in a Northern Hemisphere anticyclone.

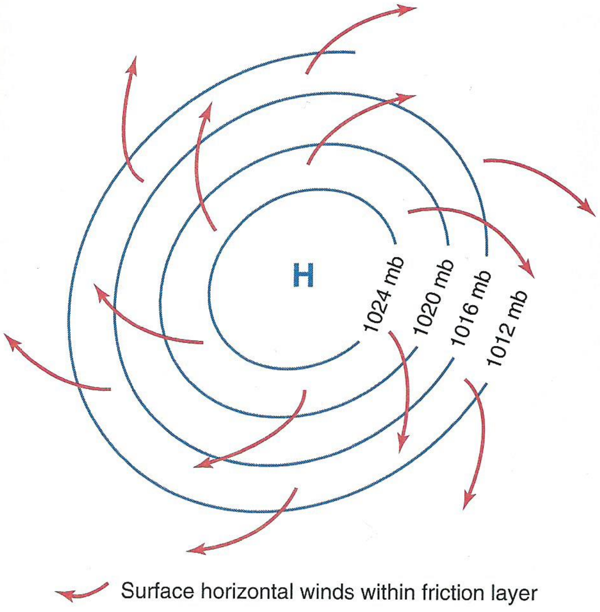
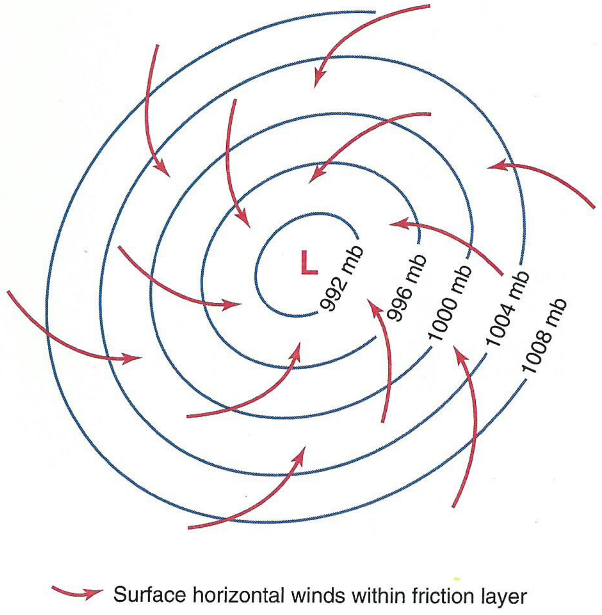


FIGURE 9.18
Viewed from above, surface winds blow counterclockwise and inward in a Northern Hemisphere cyclone.



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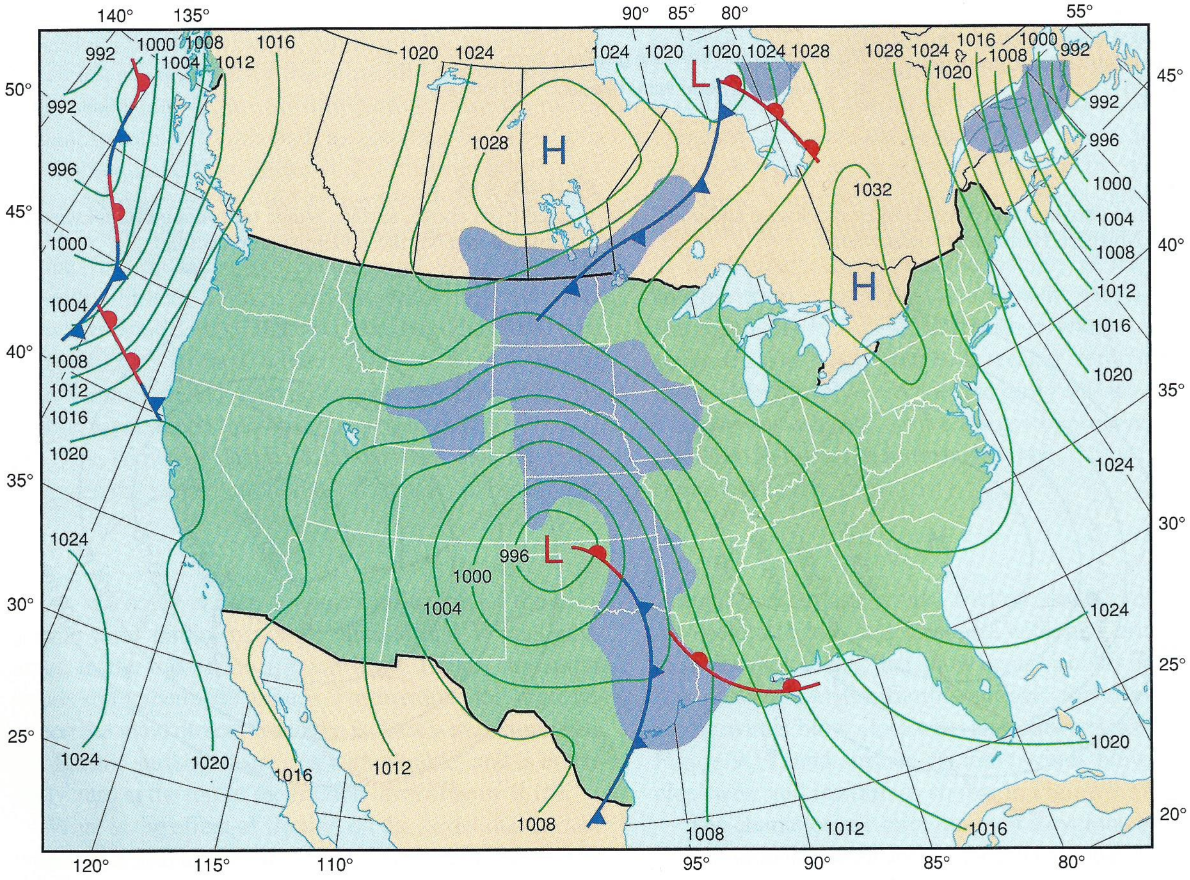
centers. If you stand with your back to the wind and then turn approximately 45 degrees to your right, the cyclone center will be located to your left. This rule is a modification of an observation first stated in 1857 by Dutch meteorologist Christopher H. D. Buys-Ballot. It must be applied with caution, however, because large-scale surface winds may favor local air circulation such as a sea breeze.

In the Southern Hemisphere, cyclonic and anticyclonic circulations are opposite their Northern Hemisphere counterparts. This contrast is due to the change in direction of the Coriolis deflection between the two hemispheres. In the Southern Hemisphere, surface winds in a cyclone blow in a clockwise and inward direction, and surface winds in an anticyclone blow in a counterclockwise and outward direction. Above the friction layer, Southern Hemisphere cyclonic winds are clockwise and parallel to isobars, and Southern Hemisphere anticyclonic winds are counterclockwise and parallel to isobars.

A glance at almost any national weather map (Figure 9.19) reveals that isobars seldom describe lengthy straight segments or circular patterns. Instead, isobars often form patterns of ridges and troughs. Nonetheless, in ridges and troughs, winds tend to parallel isobars above the friction layer and cross isobars toward low pressure within the friction layer.

An additional consideration in analyzing isobaric patterns for wind is the spacing of isobars. As noted earlier, the steeper the air pressure gradient, the faster is the wind. Where isobars are closely spaced, the geostrophic and gradient winds are relatively strong. Where isobars are widely spaced, these winds are weak. The same rule applies to surface winds.

FIGURE 9.19
Note how isobars on a typical surface weather map describe trough and ridge patterns. Shaded areas indicate precipitation. [NOAA weather map]



THERMAL WIND

The horizontal wind usually continues to increase in speed with altitude well above the friction layer. This is because the horizontal pressure gradient strengthens with increasing altitude, giving rise to a component of the geostrophic (or gradient) wind known as the **thermal wind.** The thermal wind is so named because the change in horizontal pressure gradient with altitude is due to a horizontal gradient in temperature.

Suppose that a cold air mass is situated next to a warm air mass with a *front* defining the boundary or transition between them. Even if at the Earth's surface the horizontal pressure gradient is weak across the front, the horizontal pressure gradient will steepen with increasing altitude. This is because cold air is denser than warm air, so air pressure falls faster with altitude in cold air than in warm air. The same pressure drop requires a greater ascent in warm air than in cold air. Hence, at a specific altitude, air pressure is greater in the warm air than in the cold air. This horizontal pressure gradient accelerates air away from warm air and toward cold air. Deflection of the wind by the Coriolis effect means that the thermal wind blows parallel to isotherms with cold air to the left of the direction of air motion.

In summary, the thermal wind blows parallel to a front with the cold air to the left of the direction of air movement. The thermal wind adds to the low-level geostrophic (or gradient) wind so that the wind speed usually continues to increase with altitude above the friction layer and up to the tropopause. As is demonstrated in Chapter 10, in some areas, horizontal winds near the tropopause are exceptionally strong and constitute a jet stream.

Continuity of Wind

Air is a continuous fluid, and this continuity implies a link between the horizontal and vertical components of the wind. For example, surface winds are forced to follow the undulating topography of the Earth's surface and ascend hills and descend into valleys. In addition, uplift occurs along frontal surfaces as one air mass moves horizontally and either overrides or pushes under another air mass (Chapter 6). Having examined the horizontal circulation in anticyclones and cyclones, we can identify other important connections between the horizontal and vertical components of the wind.

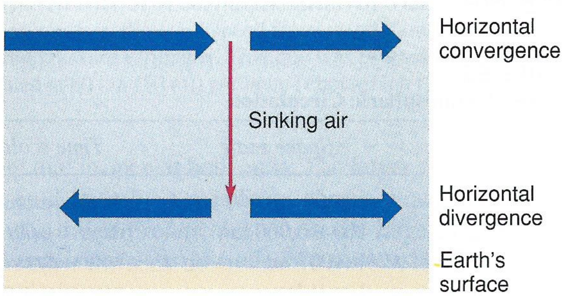
As noted earlier, surface winds in a Northern Hemisphere anticyclone spiral clockwise and outward. Consequently, the horizontal surface winds diverge away from the center of the high. A vacuum does not develop at the center, however, because air slowly descends toward the Earth's surface and replaces the air that is diverging. Air is driven downward by air that converges toward the high center aloft (Figure 9.20). Recall that adiabatic compression raises the temperature and lowers the relative humidity of descending air. Skies therefore tend to be fair within anticyclones, and anticyclones are appropriately described as *fair weather* systems.

Furthermore, within an anticyclone, the horizontal air pressure gradient is typically very weak over a broad area around the center of the system. The resulting light or calm winds coupled with clear night skies favor intense radiational cooling. Hence, the ground and the air adjacent to the ground may be chilled to the point that dew, frost, or radiation fog develops.

Surface winds in a Northern Hemisphere cyclone spiral counterclockwise and inward. Surface winds therefore converge toward the center of a low. Air does not simply pile up at the center; rather, the air ascends in response to diverging air aloft (Figure 9.21). Recall that adiabatic expansion lowers the temperature and increases the relative humidity of ascending air. Clouds and precipitation may eventually develop, so that cyclones are typically *stormy weather* systems.

Continuity of the wind also means that vertical motion can be induced by downwind changes in frictional resistance. The rougher the Earth's surface, the more resistance it offers to horizontal winds. When the horizontal wind blows from a rough surface to a relatively

FIGURE 9.20
In this idealized vertical cross section of an anticyclone, air converges aloft, sinks, and diverges at the Earth's surface.



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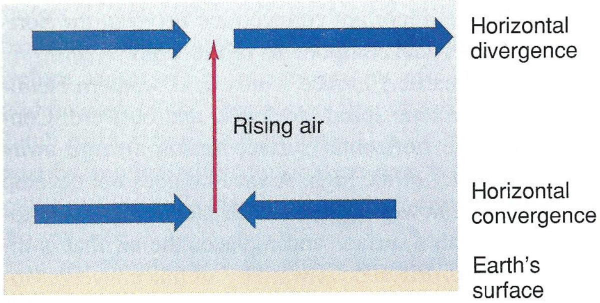

FIGURE 9.21
In this idealized vertical cross section of a cyclone, air converges at the Earth's surface, rises, and diverges aloft.

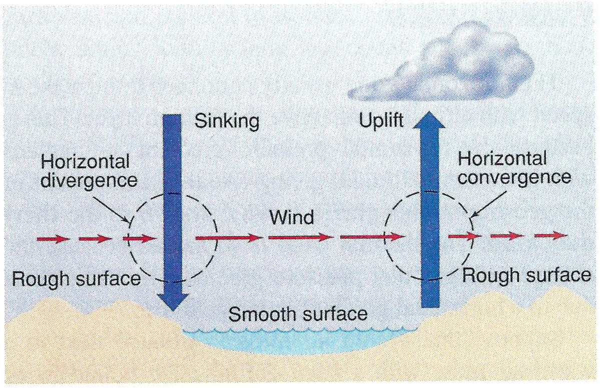

FIGURE 9.22
Surface winds undergo horizontal divergence when blowing from a rough to a smooth surface, and horizontal convergence when blowing from a smooth to a rough surface. Horizontal divergence causes air to sink and horizontal convergence causes air to rise.

smooth surface—as when it blows from land to sea—the wind accelerates. As Figure 9.22 shows, this acceleration causes the wind to diverge (stretch), thereby inducing downward motion of air. In contrast, when the horizontal wind blows from a smooth to a rough surface, the wind slows and converges (piles up), thereby inducing upward air motion. This is one reason why, along a coastline, cumuliform clouds (cumulus) tend to develop with an onshore wind (from sea to land) and tend to dissipate with an offshore wind (from land to sea).

Scales of Weather Systems

Although the atmosphere is a continuous fluid, for convenience of study we subdivide atmospheric circulation into discrete weather systems that operate at various spatial and temporal scales (Table 9.2). The large-scale wind belts encircling the planet (polar easterlies, westerlies, and trade winds) are global or **planetary-scale systems. Synoptic-scale systems** are continental or oceanic in scale; migrating cyclones, hurricanes, and air masses are examples. **Mesoscale systems** include thunderstorms and sea and lake breezes—phenomena that are so small that they may influence the weather in only a portion of a large city. A circulation system covering a very small area—a tornado, for example—represents the smallest spatial subdivision of atmospheric motion, **microscale systems.**

TABLE 9.2
Scales of Atmospheric Circulation

Circulation	*Space scale*	*Time scale*
Planetary-scale	10,000–40,000 km	weeks–months
Synoptic-scale	100–10,000 km	days–weeks
Mesoscale	1–100 km	hours–days
Microscale	1 m–1 km	seconds–hours

Circulation systems not only differ in spatial scale, they also contrast in life expectancy. Thus, patterns in the planetary-scale circulation may persist for weeks or even months. Synoptic-scale systems typically last for several days to a week or so. Mesoscale systems usually complete their life cycles in a matter of hours to perhaps a day, whereas microscale systems persist for minutes or less.

There are other differences among the various scales of atmospheric circulation. At the micro- and mesoscale, vertical velocities may be comparable in magnitude to horizontal velocities. At the synoptic and planetary scales, however, horizontal winds are considerably stronger than vertical flow. Furthermore, at the micro- and mesoscale, the Coriolis effect is usually negligibly small. In contrast, the Coriolis effect is very important in synoptic- and planetary-scale circulation systems.

Each smaller scale weather system is part of, and dependent on, larger scale atmospheric circulation. That is, the various scales of atmospheric motion form a kind of hierarchy. For example, extreme nocturnal radiational cooling requires a synoptic weather pattern that favors clear skies and light or calm winds. At the microscale, such weather conditions may be accompanied by frost formation or radiation fog.

In Chapters 10 through 15, we examine weather systems of all scales, starting with the planetary scale.

Wind Pressure

As wind speed increases, the potential for wind damage to trees, buildings, and other structures also increases. In this regard, the concept of wind pressure is useful. **Wind pressure** is defined as the force per unit area produced by the wind on an object in its path. Wind pressure is directly proportional to the square of the wind speed (v^2). Hence, each doubling of wind speed increases the wind pressure by a factor of four ($2^2 = 4$).

In designing buildings to withstand high winds, engineers and architects consult the climatic record to determine the probability of potentially destructive winds. Because wind speed increases with altitude, a building's upper stories are subject to greater wind pressure than its lower stories. A building's exposure is also an important consideration in planning for wind pressure. For example, a building situated on a coastal plain and facing the open ocean is exposed to stronger winds than a building situated among many other buildings within a congested city center.

In the reverse application of this approach, wind pressure, and hence wind speed, can be reconstructed from storm damage (Figure 9.23). In this way, for example, meteorologists have been able to estimate wind speeds in tornadoes based on the damage they produce. Such reconstructions are necessary because tornadic winds are so strong that they destroy conventional weather instruments.

FIGURE 9.23
In the absence of direct measurements of wind, it is possible to reconstruct the wind speed from storm damage based on wind pressure estimates. This tree was twisted by winds estimated at 80 km (50 mi) per hour. [Photograph by J. M. Moran]

Wind Measurement

Meteorologists are interested in monitoring both the speed and direction of wind. Most wind-monitoring instruments are designed to measure only the horizontal component of the wind because it is usually considerably stronger than the vertical component. For some specialized research purposes, very sensitive instruments are available that measure vertical wind speeds or a combination of vertical and horizontal wind components.

A traditional **wind vane,** like the one shown in Figure 9.24, consists of a free-swinging horizontal shaft with a vertical plate at one end and a counterweight (arrow) at the other end. The counterweight always points directly into the wind. Another design is the airport **wind sock,** which consists of a cone-shaped cloth bag that is open at both ends. The larger end of the sock is held open by a metal ring that is attached to a pole and is free to rotate. Air enters the larger opening and stretches the sock downwind (Figure 9.25).

Wind direction is always designated as *the direction from which* the wind blows. For example, a wind

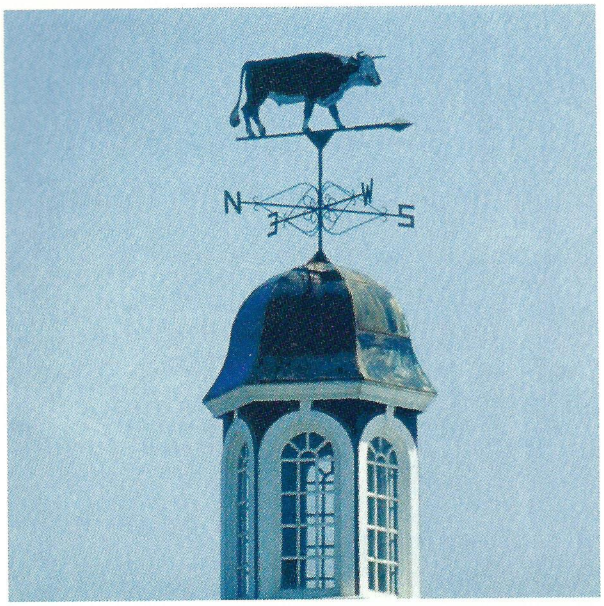

FIGURE 9.24
Horizontal wind direction is monitored by a wind vane. The instrument's arm points in the direction from which the wind blows. [Photograph by J. M. Moran]

blowing from the east toward the west is described as an *east* wind and a wind blowing from the northwest to the southeast is a *northwest* wind. A wind vane may be linked electronically or mechanically to a dial that is calibrated to read in points of the compass or in degrees. Measured clockwise from true north, an east wind is specified as 90 degrees, a south wind as 180 degrees, a west wind as 270 degrees, and a north wind as 360 degrees. The wind is recorded as 0 degrees only under calm conditions.

Wind speed can be estimated by observing the wind's effect on lake or ocean surfaces or on land-based flexible objects such as trees. Such observations are the basis of the **Beaufort scale,** which is a graduated sequence of wind strength ranging from 0 for calm conditions to 12 for hurricane-strength winds (Table 9.3). The scale bears the name of Sir Francis Beaufort, who developed it in the early 1800s while a ship commander in the British Navy. Beaufort's goal was to standardize terms used by sailors in describing the state of the sea under various wind conditions. In 1838, after some revision, the British Navy adopted the Beaufort scale, and in 1853, it was sanctioned for international use by seafarers. Later, when the scale was extended from sea to land, it was necessary to develop wind speed equivalents for each Beaufort number; this was done in 1926. The scale is still used today. In fact, some modern-day mariners prefer Beaufort numbers to onboard instrument measurements of wind speed.

FIGURE 9.25
An airport wind sock gives wind direction and a general indication of wind speed. The sock points downwind. [Photograph by J. M. Moran]

A **cup anemometer** consists of 3 or 4 open hemispheric cups mounted on a vertical shaft, such as the one shown in Figure 9.26. At least one open cup faces the wind at any time. Rotation of the array of cups is calibrated by computer to read in meters per second, kilometers per hour, or knots.* Several other types of anemometers are available, including the very sensitive **hotwire anemometer.** In this instrument, the wind blows past a heated wire, or wires, and the heat lost to moving air is calibrated in terms of wind speed.

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*One knot = 1 nautical mile (1.85 km) per hour; 1 knot = 0.51 meter per second; 1 mile per hour = 0.44 meter per second.

TABLE 9.3
Beaufort Scale of Wind Force

Beaufort number	*General description*	*Land and sea observations for estimating wind speeds*	*Wind speed 10 m above ground (km/hr)*
0	Calm	Smoke rises vertically. Sea like mirror.	Less than 1
1	Light air	Smoke, but not wind vane, shows direction of wind. Slight ripples at sea.	1–5
2	Light breeze	Wind felt on face, leaves rustle, wind vanes move. Small, short wavelets.	6–11
3	Gentle breeze	Leaves and small twigs moving constantly, small flags extended. Large wavelets, scattered whitecaps.	12–19
4	Moderate breeze	Dust and loose paper raised, small branches moved. Small waves, frequent whitecaps.	20–28
5	Fresh breeze	Small leafy trees swayed. Moderate waves.	29–38
6	Strong breeze	Large branches in motion, whistling heard in utility wires. Large waves, some spray.	39–49
7	Near gale	Whole trees in motion. White foam from breaking waves.	50–61
8	Gale	Twigs break off trees. Moderately high waves of great length.	62–74
9	Strong gale	Slight structural damage occurs. Crests of waves begin to roll over. Spray may impede visibility.	75–88
10	Storm	Trees uprooted, considerable structural damage. Sea white with foam, heavy tumbling of sea.	89–102
11	Violent storm	Very rare; widespread damage. Unusually high waves.	103–118
12	Hurricane	Very rare; much foam and spray greatly reduce visibility.	119 and over

Recording a continuous trace of wind speed and direction is sometimes informative. Wind vanes and anemometers can be linked to pens that record on a paper chart that is attached to a clock-driven drum. As shown in Figure 9.27, the trace indicates a considerable variation in both wind direction and wind speed with time. The spectrum of wind gusts and lulls indicate turbulence. More commonly today, the output from wind vanes and anemometers is recorded by computer or on magnetic tape.

Ideally, a wind vane or anemometer system should be mounted on a tower so that the instruments monitor horizontal winds 10 m (33 ft) above the ground. It is best to avoid rooftop locations because winds tend to accelerate over buildings. In addition, the system should be sited well away from (1) structures that might shelter the instruments and (2) any obstacles that might channel (and thus accelerate) the wind.

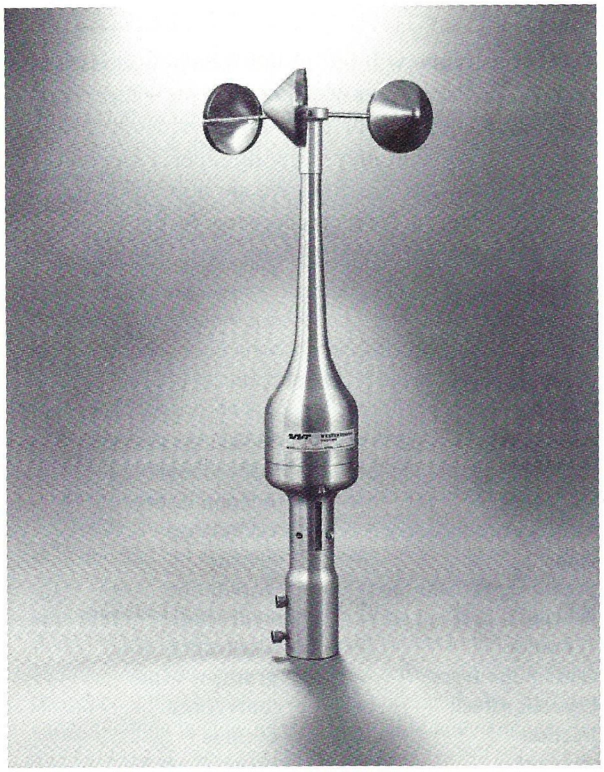

FIGURE 9.26
Wind speed is measured by a cup anemometer. The faster the wind speed, the faster the cups spin. [Courtesy of Qualimetrics]

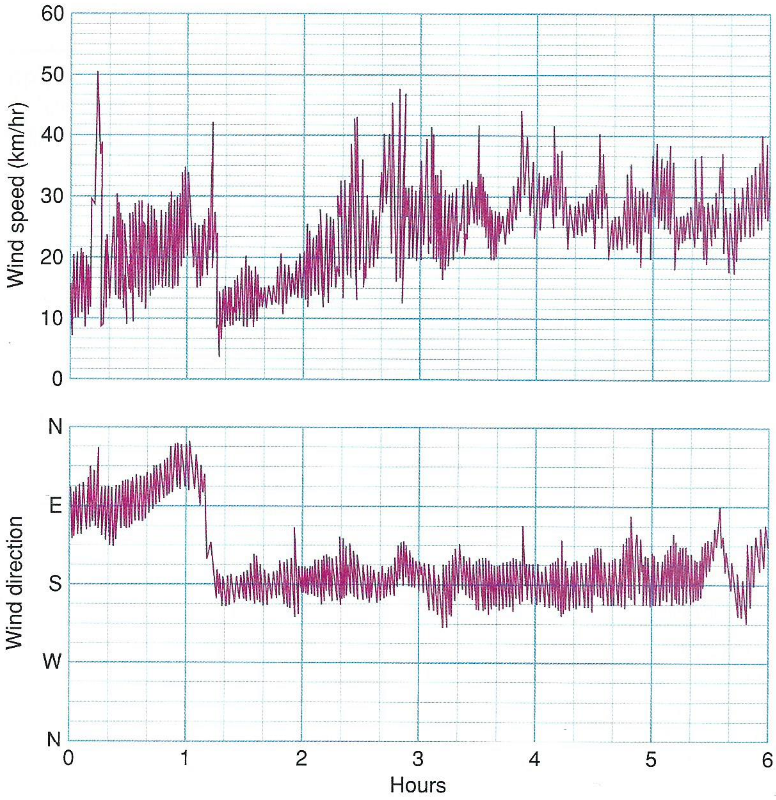

FIGURE 9.27 Continuous trace of time variations in wind speed and wind direction over a six-hour period.

Conclusions

In this book so far, we have seen that unequal rates of radiational heating and cooling in the Earth–atmosphere system give rise to gradients in temperature. In response to those gradients, the atmosphere circulates and thereby, heat energy is converted to kinetic energy. In this chapter, we have examined the various forces that initiate and shape atmospheric circulation (the wind). Note that the pressure gradient force and gravity would exist even if air were not in motion, and the other forces (centripetal, Coriolis, and friction) come into play only after air is in motion.

From everyday experience, we are aware of the end product of the forces described in this chapter, namely, the wind, but we are not readily aware of the individual forces. In learning about atmospheric forces and how they interact, we see that each force is bound by certain constraints. For example, friction is important only in the lower troposphere, and the Coriolis effect always shifts the wind to the right in the Northern Hemisphere. In the following chapters, our awareness of these and other constraints will aid in understanding the characteristics of the various weather systems—for example, why hurricanes do not form at the equator, and why winds in a tornado may blow in either a clockwise or counterclockwise direction. We are now ready to begin our discussion of the characteristics of the various circulation systems, beginning with those operating at the planetary scale.

Key Terms

wind
Newton's second law of motion
air pressure gradient
isobars
pressure gradient force
Newton's first law of motion
centripetal force
Coriolis effect
friction
viscosity
molecular viscosity
eddy viscosity
friction layer
turbulence
gravity
gravitation
hydrostatic equilibrium
geostrophic wind
gradient wind
anticyclone
cyclone
thermal wind

WEATHER FACT

The Windiest Place on Earth

On 12 April 1934, the weather station at the 1910-m (6262-ft) summit of Mount Washington, New Hampshire, recorded a peak wind gust of 373 km (231 mi) per hour. This is the highest wind speed ever recorded.* That same day, the wind averaged 303 km (188 mi) per hour over a span of 5 minutes. At 57 km (35 mi) per hour, the average annual wind speed on Mount Washington is the greatest of any location in the United States. Average monthly wind speed ranges from about 40 km (25 mi) per hour in July to about 73 km (45 mi) per hour in January. Winds in excess of hurricane force are common in winter.

Elevation and exposure factor into the windiness of Mount Washington and other mountaintops. As noted elsewhere in this chapter, the *thermal wind* causes the westerlies to strengthen with increasing altitude. In addition, as air flows up and over a mountain peak, it is squeezed between the summit and overlying air layers. This constriction accelerates the wind. The prevailing westerlies are normally relatively strong over New England and the constricted flow over Mount Washington accelerates them to extreme magnitudes.

*There is no doubt that stronger winds occur in violent tornadoes, but so far they have not been recorded by anemometers.

planetary-scale systems
synoptic-scale systems
mesoscale systems
microscale systems
wind pressure
wind vane
wind sock
Beaufort scale
cup anemometer
hotwire anemometer

Summary Statements

☐ Wind is the movement of air measured relative to the Earth's surface. Wind is the consequence of interactions of the pressure gradient force, the Coriolis effect, friction, and gravity. The centripetal force is the resultant of other forces.

☐ The pressure gradient force initiates air motion and arises in part from spatial variations in air temperature and, to a lesser extent, water vapor concentration. In response to gradients in pressure, air accelerates from areas of relatively high pressure toward areas of relatively low pressure.

☐ Wind is a vector quantity; that is, it has both direction and magnitude. Hence, an acceleration of the wind may consist of a change in speed or direction or both.

☐ The centripetal force is the imbalance in actual forces that operates whenever the wind follows a curved path. The centripetal force is responsible for a change in direction of the wind and not a change in speed.

☐ The Coriolis effect arises from the Earth's rotation on its axis. It deflects the wind to the right of its initial direction in the Northern Hemisphere and to the left in the Southern Hemisphere. The deflective force is zero at the equator and increases with latitude to a maximum at the poles. The Coriolis effect is important only in large-scale (planetary- and synoptic-scale) circulation systems.

☐ Friction affects horizontal winds blowing within about 1 km (0.62 mi) of the Earth's surface. Obstacles on the Earth's surface slow the wind by breaking it into turbulent eddies.

☐ Gravity always accelerates objects downward and perpendicular to the Earth's surface. Gravity is important for vertical motion of air.

☐ Hydrostatic equilibrium is the balance between the upward-directed pressure gradient force and the downward-directed force of gravity. Slight deviations in hydrostatic equilibrium cause air to accelerate upward or downward.

☐ The geostrophic wind is an unaccelerated, horizontal wind that blows in a straight path parallel to isobars at altitudes above the friction layer. The geostrophic wind results from a balance between the horizontal pressure gradient force and the Coriolis effect.

☐ The gradient wind is a horizontal wind that parallels curved isobars at altitudes above the friction layer. The centripetal force operates in the gradient wind and is the result of an imbalance between the horizontal pressure gradient force and the Coriolis effect. In the Northern Hemisphere, the gradient wind blows clockwise in anticyclones and counterclockwise in cyclones.

☐ In large-scale (synoptic- and planetary-scale) circulation systems, friction slows the wind and interacts with the Coriolis effect to shift the wind direction across isobars and toward low pressure.

☐ Within the friction layer, horizontal winds blow clockwise and outward in Northern Hemisphere anticyclones and counterclockwise and inward in Northern Hemisphere cyclones.

☐ In an anticyclone, horizontal divergence of surface winds causes descending air. Hence, an anticyclone is a fair-weather

MATHEMATICAL NOTE

Geostrophic and Gradient Winds

The geostrophic wind is the result of a balance between the horizontal pressure gradient force and the Coriolis effect. Here we examine the geostrophic motion of a unit mass of air (1 g, for example). The acceleration imparted to this air parcel by a horizontal pressure gradient is given by

$$\frac{1}{\rho}\left(\frac{\Delta P}{\Delta N}\right)$$

where ρ is the density of the air and ΔP is the change in air pressure over a horizontal distance, ΔN, measured perpendicular to isobars. (Pressure gradients are always measured perpendicular to isobars.)

The acceleration imparted to the air parcel by the Coriolis effect is given by

$$(2\Omega \sin \phi)\, v$$

where v is the speed of the air parcel, Ω is the angular velocity of the Earth as it rotates on its axis ($\Omega = 7.29 \times 10^{-5}$ radian per second), and ϕ is the latitude.

For the geostrophic wind,

$$\frac{1}{\rho}\left(\frac{\Delta P}{\Delta N}\right) = (2\Omega \sin \phi)\, v$$

Solving for v, the speed of the geostrophic wind, we have

$$v = \frac{1}{(2\Omega \sin \phi)\rho}\left(\frac{\Delta P}{\Delta N}\right)$$

The geostrophic wind thus strengthens with an increasing pressure gradient (that is, closer spacing of isobars) and decreasing latitude. Recall that the geostrophic wind blows parallel to isobars and at right angles to the pressure gradient force and Coriolis effect (which are directed one opposite the other).

The gradient wind results from a slight imbalance between the horizontal pressure gradient force and the Coriolis effect. This imbalance is realized as a centripetal force. Again, assume that we are examining the motion of a unit mass of air. The acceleration imparted to the air parcel by the centripetal force is

$$v^2/r$$

where v is the speed of the air parcel, and r is the radius of curvature of the path described by the air parcel. The centripetal force is the force that constrains the air parcel to a curved trajectory. This force is strong where the curvature is sharp (small r) and weak where the curvature is gradual (large r). The three forces (pressure gradient, Coriolis, and centripetal) interact such that

$$\frac{v^2}{r} + (2\Omega \sin \phi)\, v - \frac{1}{\rho}\left(\frac{\Delta P}{\Delta N}\right) = 0$$

The net force, which is the centripetal force, operates on the gradient wind only to change the direction of the wind as it follows a curved path, and not to change the wind's speed. We could solve the above equation for v to determine the speed of the gradient wind for some radius of curvature, latitude, and horizontal pressure gradient (measured perpendicular to isobars). Note that for wind blowing in a straight line,

$$v^2/r = 0$$

and the equation reduces to that presented earlier for the geostrophic wind.

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system. In a cyclone, horizontal convergence of surface winds causes ascending air. Hence, a cyclone is a stormy weather system.

☐ Along a coastline, offshore winds undergo horizontal divergence (including descending air), whereas onshore winds undergo horizontal convergence (including ascending air).

☐ Atmospheric circulation is divided into four spatial/temporal scales: planetary, synoptic, mesoscale, and microscale.

Review Questions

1. Provide a definition for *wind*. Compare the magnitude of the vertical wind versus the horizontal wind in large-scale weather systems. Do the same for meso- and microscale systems.
2. What causes horizontal gradients in air pressure?
3. What is the relationship between horizontal air pressure gradients and wind speed and direction?

4. Why must a centripetal force operate whenever the wind follows a curved trajectory?
5. What causes the Coriolis effect and why does its magnitude increase with increasing latitude? Why is the Coriolis deflection important only in large-scale (planetary- and synoptic-scale) weather systems?
6. Why does the Coriolis deflection reverse between the Northern and Southern hemispheres?
7. Distinguish between *molecular viscosity* and *eddy viscosity.*
8. How does the horizontal wind speed change with altitude within the friction layer?
9. State Newton's first and second laws of motion.
10. Provide an example of how gravity influences the motion of air.
11. Define *hydrostatic equilibrium.* Does hydrostatic equilibrium imply that there can be no ascent or descent of air within the atmosphere? Explain your response.
12. Distinguish between the geostrophic wind and the gradient wind.
13. What is responsible for centripetal forces in cyclones and anticyclones?
14. How does the roughness of the Earth's surface affect the horizontal wind direction?
15. Describe the horizontal circulation in a cyclone (a) within the friction layer and (b) above the friction layer.
16. Describe the horizontal circulation in an anticyclone (a) within the friction layer and (b) above the friction layer.
17. Why is fair weather usually associated with anticyclones and stormy weather with cyclones?
18. Horizontal air pressure gradients are usually weak over a wide area around the center of an anticyclone. What does this imply about the weather at the center of an anticyclone?
19. How do downwind changes in surface roughness (frictional resistance) induce divergence and convergence of the horizontal wind? Provide specific examples.
20. Present examples of planetary-scale, synoptic-scale, mesoscale, and microscale circulation systems.

Questions for Critical Thinking

1. Suppose that the magnitude of the horizontal pressure gradient changes with altitude. How would this affect the horizontal wind?
2. In view of Newton's first law of motion, is gradient wind the consequence of *balanced* forces? Explain your response.
3. How does the horizontal pressure gradient compare in magnitude to the vertical pressure gradient?
4. Predict how wind direction and speed at your locality change as a cyclone approaches. Do the same for an approaching anticyclone.
5. Suppose that a synoptic-scale cyclone is centered over central Kansas. Describe the air mass advection to the southeast, west, and northeast of the storm center.

Selected Readings

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