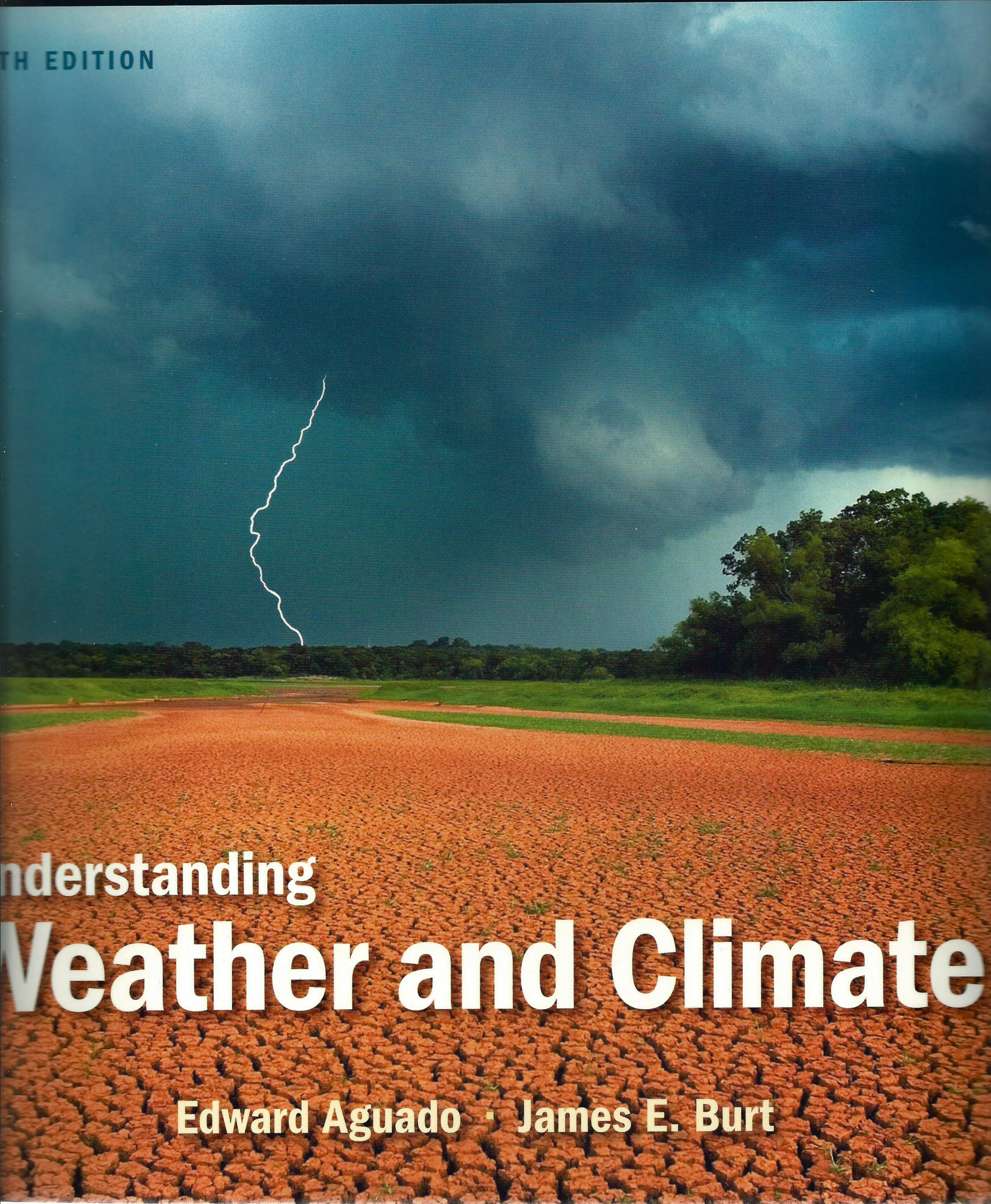

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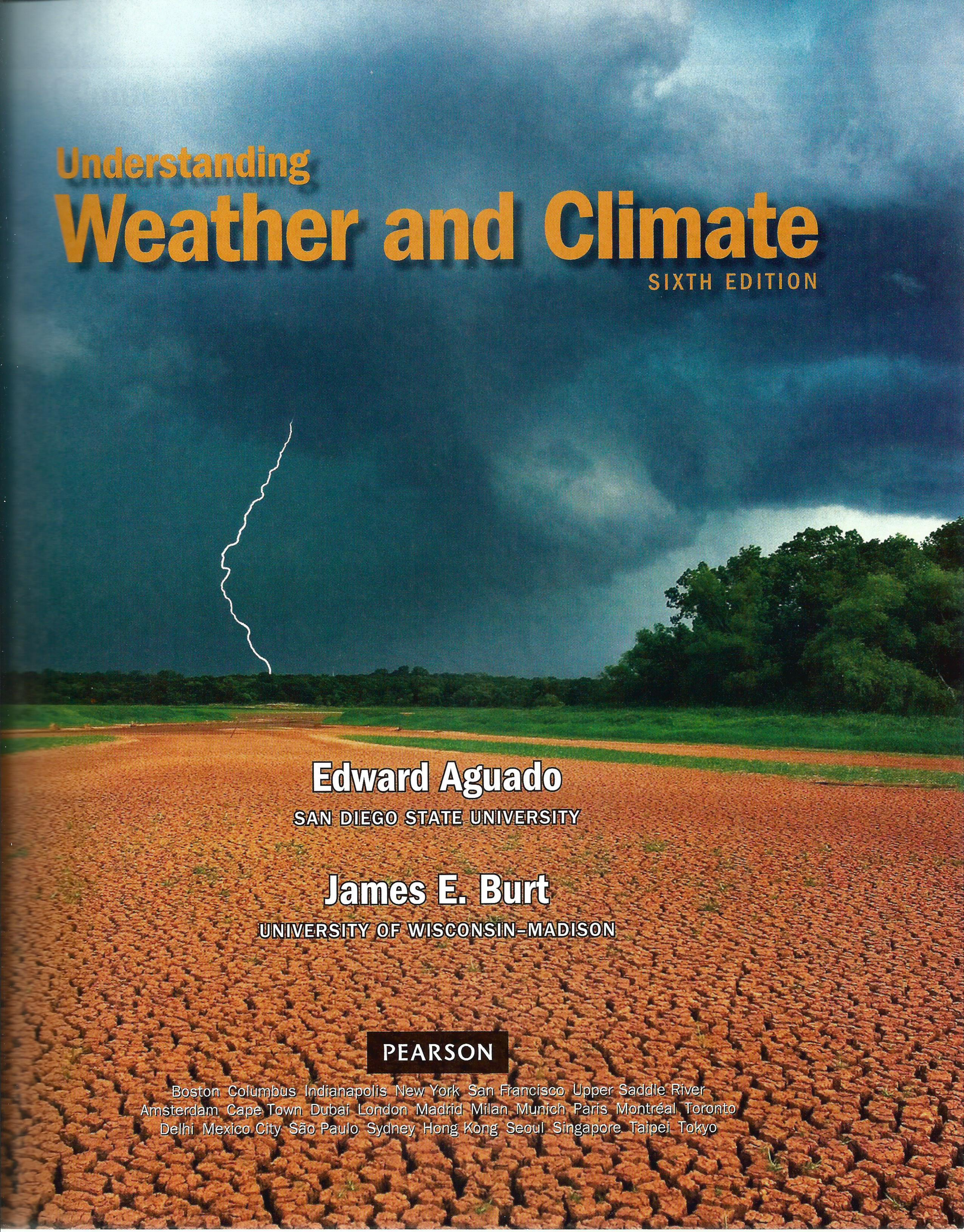
Edward Aguado · James E. Burt



Understanding Weather and Climate

SIXTH EDITION

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UNIVERSITY OF WISCONSIN–MADISON

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ISBN 10: **0-321-76963-5;** ISBN 13: **978-0-321-76963-3** [Student Edition]
ISBN 10: **0-321-77323-3;** ISBN 13: **978-0-321-77323-4** [Instructor's Review Copy]

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7

Precipitation Processes

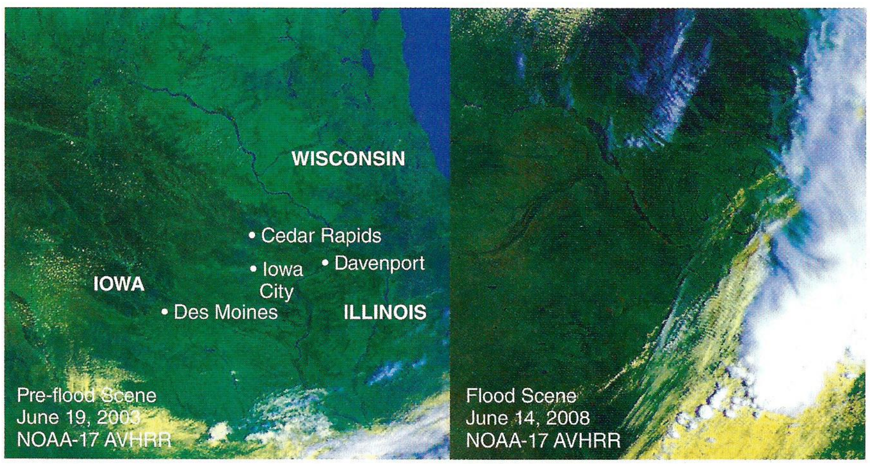
LEARNING OUTCOMES

After reading this chapter, you should be able to:

- Describe the processes involved in the growth of cloud droplets.
- Describe the distribution of precipitation and explain how different types of precipitation form.
- Describe how precipitation is measured.
- Summarize efforts to induce precipitation through cloud seeding.

During the spring of 1993 residents of the Midwestern United States witnessed flooding on an unprecedented scale. At St. Louis, Missouri, the Mississippi River crested at 6 m (19.5 ft) above flood stage. The river, normally about 800 m (0.5 mi) wide near St. Joseph, Missouri, stretched out to as much as 10 km (6 mi), putting nearly half of St. Charles County under water. At Kansas City, Missouri, the Missouri River rose 6.7 m (22 ft) above its banks. Across the Midwest, tens of thousands of homes were damaged or destroyed by the flooding, as entire neighborhoods and 77 small towns ended up under water. The flooding even brought its share of irony: Des Moines, Iowa, was without potable water for 12 days because of contaminated floodwaters. Forty-year-old Jacki Meek of suburban St. Louis probably spoke for all of the 85,000 people who had to evacuate their homes: "I feel about 65 right now. I see my house on the news, and I just cry."

But what many would have thought would be a once-in-a-lifetime event was repeated 15 years later, when another round of extensive flooding soaked the Midwest in 2008. A series of heavy rains hit the region in early June, including a few exceptionally strong ones that brought more than 22 cm (9 in.) of rain in a 2-day period. Gays Mills, Wisconsin, which had been inundated by flood waters from the Kickapoo River for the second time in 10 months, was forced to consider relocating the town on higher ground farther up the floodplain to avoid similar events in the future.



▲ **FIGURE 7–1** Flooding occurred over a large portion of the upper Midwest in June 2008. These satellite images show the contrast between the normal situation and that which existed during the peak of the flooding. Note the increased width of the rivers in 2008, especially the Cedar River west of Iowa City and Cedar Rapids, Iowa, and the amount of saturated ground.

Indiana, Michigan, Illinois, and Missouri were beset by record-breaking floods from exceptionally heavy rain. On June 7, Edinburgh, Indiana, received 27.2 cm (10.71 in.) of rain, the highest amount ever recorded in a single day for the entire state. But no state was hit harder than Iowa, where 83 of its 99 counties were declared disaster areas, more than 8 percent of the corn and soybean acreage were under water (Figure 7–1), and damage was estimated at about $1.5 billion. Many towns and cities fought rising water with sandbags, but often unsuccessfully, as in Cedar Rapids (Figure 7–2).

◄ Record flooding inundates homes in Cedar Rapids, Iowa, in June 2008.

▲ **FIGURE 7–2** Flood damage at Cedar Rapids, Iowa, June 2008.

Overall, how did the 1993 and 2008 floods compare? They were similar in many respects. The 1993 flooding covered a larger area and maintained excessive river levels for a longer period of time. The 2008 floods were the result of more intense rainfall events that occurred over a shorter time period and had faster falling river levels after the peak of the flooding.

Rain and other forms of precipitation are a fact of life for everybody, although usually they are of far less consequence than the floods of 1993 and 2008. The desire to know what causes precipitation may have been one reason you picked up this book. In Chapter 6 you learned about the processes that lead to the formation of all clouds, precipitating or nonprecipitating. In this chapter we explain the processes by which nonprecipitating cloud droplets and ice crystals grow large enough to fall as precipitation.

Growth of Cloud Droplets

Acting alone, gravity would accelerate all objects toward the surface. But gravity is not the only force acting on a falling object; at the same time, the air exerts an opposing resistance or **drag**. As speed increases, so does resistance, until its force equals that of gravity and the acceleration ceases. The object falls, but at a constant speed, its **terminal velocity.** More than anything else, terminal velocity depends on size, with small objects falling much more slowly than large objects. (We examine details of the relationship between size and terminal velocity in *Box 7–1, Physical Principles: Why Cloud Droplets Don't Fall.*) Cloud droplets fall slowly because they are so tiny. Their small size is largely explained by the fact that condensation nuclei are very abundant; thus, cloud water is spread across numerous small droplets rather than being concentrated in fewer large drops. With their small size, cloud droplets initially have extremely low terminal velocities, making it impossible for them to reach the surface.

This effect is apparent in Figure 7–3, which shows terminal velocities for various cloud constituents. The smallest are the condensation nuclei, on which liquid droplets form. (For the sake of simplicity, the figure applies only to clouds consisting of liquid water alone, without ice crystals.) Condensation nuclei are so small that they fall at an imperceptibly slow rate. Larger cloud droplets (but not falling as precipitation) typically range from about 10 μm to about 50 μm in radius (recall that 1 μm is one-millionth of a meter). These have fall speeds ranging from about 1 cm/sec (0.02 mph) to about 25 cm/sec (0.5 mph). By way of contrast, the much larger raindrops shown in the figure fall at 650 cm/sec, about 25 times faster.

Raindrops fall to the surface when they become large enough that gravity overcomes the effect of updrafts. How large is large enough? In terms of radius, raindrops are about 100 times bigger than typical cloud droplets. But in terms of volume or mass of water, raindrops are larger than cloud drops by a factor of a million, rather than just 100. The difference arises because volume for a sphere is proportional to the cube of the radius. If the radius is 100 times larger, the volume is 100 × 100 × 100 (1 million) times larger. Raindrops are not truly spherical, but the principle holds: Precipitation particles are vastly more massive than cloud drops. Although we do not think of clouds yielding massive falling objects, they certainly do, at least from the point of view of a cloud droplet. In the paragraphs that follow, we outline the processes that give rise to these "massive" falling objects.

7–1 PHYSICAL PRINCIPLES

Why Cloud Droplets Don't Fall

You are probably familiar with the legendary, late-sixteenth-century experiment of Galileo Galilei, who dropped two objects—a light one and a heavy one—off the Leaning Tower of Pisa. The objects, being subjected to the same gravitational acceleration, hit the ground at nearly the same time. Galileo's demonstration may seem inconsistent with our everyday experience, as an ant would surely take longer than a golf ball to fall from the top of a tall building. It is also at odds with our claim that small droplets fall slowly. The solution must be that a force besides gravity acts on falling objects: It is wind resistance, or drag. By examining how these two forces work together, we will gain some insight into why cloud droplets do not fall. To keep the discussion simple, we will assume spherical droplets throughout—using more realistic shapes would not change our conclusions.

Newton's second law tells us that if a net force is applied to a mass, it will undergo an acceleration (or change in velocity through time). For a given mass, the acceleration is directly proportional to the net force. In equation form, the law is given as

$$\text{net force} = \text{mass} \times \text{acceleration}$$

Notice that Newton's second law says that we must consider the net force, the result of all the forces acting on the object. As far as a falling droplet is concerned, there is the downward gravitational force, which is opposed by the force of wind resistance (drag). A droplet suddenly released in the atmosphere falls at increasing speed, but not indefinitely. Eventually the force of drag (F_d) balances the force of gravity (F_g), resulting in no net force:

$$\text{net force} = F_g - F_d = 0$$

With no net force, there is no acceleration, and the droplet falls at its terminal velocity. How fast does it fall? To answer that, we need to know something about the magnitude of the two forces.

Force of Gravity

The force of gravity is equal to mass times the acceleration of gravity, g. Whenever we step on a scale, we measure this force. For a liquid droplet presumed to be spherical, mass is the density of water, ρ, times the droplet volume, $4/3\pi r^3$, where r is the droplet radius. We therefore have F_g as

$$F_g = \rho\frac{4}{3}\pi r^3 g$$

Force of Drag

Drag between the droplet and surrounding air depends on the rate of fall and on the size of the droplet. Just like an automobile on a highway, a faster-moving droplet experiences greater resistance as it moves through the air. In fact, to a good approximation, the drag force increases with the square of wind speed, (v^2). So how does size influence drag?

It can be shown that the drag can be expressed as,

$$F_d = 0.5\, C_D \rho_a v^2 \pi r^2$$

where C_D is a constant referred to as the drag coefficient, and ρ_a is the density of air (about one thousandth the density of water). The value of C_D is not important here; what matters is that F_d is proportional to the square of both the fall rate (v^2) and the radius (r^2).

Terminal Velocity

For a droplet falling at terminal velocity, we have said that gravity and drag are equal. If we use v_t for terminal velocity, we get

$$F_g = F_d$$

$$g\rho\frac{4}{3}\pi r^3 = 0.5 C_D \rho_a v_t^2 \pi r^2$$

To find the terminal velocity, we rearrange and first solve for v_t^2

$$v_t^2 = (g\rho\frac{4}{3}\pi r^3)/(0.5 C_D \rho_a \pi r^2)$$

By consolidating the numerical values and constants in the above equation into a single constant, k_1, we get

$$v_t^2 = k_1 r$$

or

$$v_t = k_t\sqrt{r}$$

where $k_2 = \sqrt{k_1}$.

From this equation, we see that as droplet radius increases, so does terminal velocity. Large droplets fall faster than small droplets. What happens physically is that both F_g and F_d increase with radius, but the gravitational force increases more than drag, and a higher fall rate is therefore required to cancel F_g. Notice that as far as the droplet is concerned, falling through a still atmosphere at v_t is the same as remaining stationary in an updraft of speed v_t. Thus, the equation says that a strong updraft is needed to hold a large droplet aloft, whereas a small droplet is easily suspended.

Going back to the Leaning Tower of Pisa situation described at the beginning of this box, we can now understand why Galileo's objects fell at nearly the same speed. With large and therefore heavy objects, the gravitational forces far exceeded the drag forces throughout their short fall. With negligible drag, gravity accelerated both at nearly the same rate. If he had used objects of greatly different size, or if the objects had fallen far enough to reach their terminal velocities, differences in v_t would have emerged. The old, familiar story would be about wind resistance, and books like this would have no need for a feature on the topic.

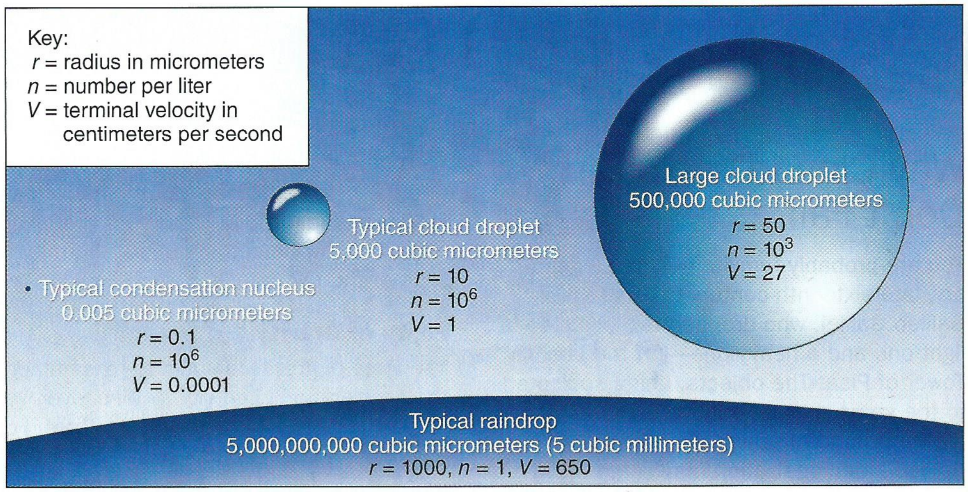
Growth by Condensation

When cloud droplets begin to form by the adiabatic cooling of ascending air, they do so on condensation nuclei. But within a few dozen meters above the lifting condensation level, all the available condensation nuclei have attracted water, and any further condensation can only occur on existing droplets.

TUTORIAL
PRECIPITATION

Use the tutorial to explore the relationship between droplet radius and terminal velocity, and to see how collision and coalescence are related to droplet size.

▸ **FIGURE 7-3** The average characteristics of cloud constituents.



Condensation can lead to rapid growth for very small water droplets, but only until they achieve radii up to about 20 μm—far smaller than necessary to fall as precipitation. Beyond this point, further growth by condensation is minimal. To understand why, recall that relatively little water vapor is available for condensation. With so many droplets competing for a limited amount of water, none can grow very large. It is clear that if growth by condensation were the only process operating, we would experience little, if any, precipitation on Earth. We should therefore think of condensation as only the starting point for rain and snow. Two other processes are responsible for further droplet growth; their relative importance depends on the temperature of clouds.

Growth in Warm Clouds

Most precipitating clouds in the tropics, and some in the middle latitudes, are **warm clouds**, those having temperatures greater than 0 °C throughout. In warm clouds, the **collision–coalescence process** causes precipitation. This process depends on the differing fall speeds of different-sized droplets.

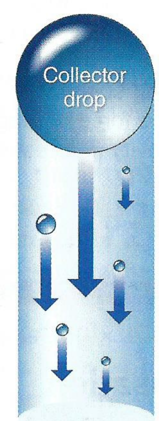
Cloud droplets come in different sizes, and therefore attain different terminal velocities. Refer to Figure 7–4 and consider what will happen when the largest droplet (the **collector drop**) falls through a warm cloud. As the collector drop falls, it overtakes some of the smaller droplets in its path because of its greater terminal velocity. This provides the opportunity for collisions and coalescence.

Collision As it falls, a collector drop collides with only some of the droplets in its path. The likelihood of a **collision** depends on both the absolute size of the collector and its size relative to the droplets below. If the collector drop is much larger than those below, the percentage of collisions (the *collision efficiency*) will be low. Figure 7–5 illustrates why. As the collector drop falls, it compresses the air in its path. The compressed air creates a small gust of wind that pushes the smaller droplets out of the way. The small gust of wind cannot push aside larger droplets, however, and the collector is able to collide with them. As a result, the collision efficiency is lower for droplets that are very much smaller than the collector drops.

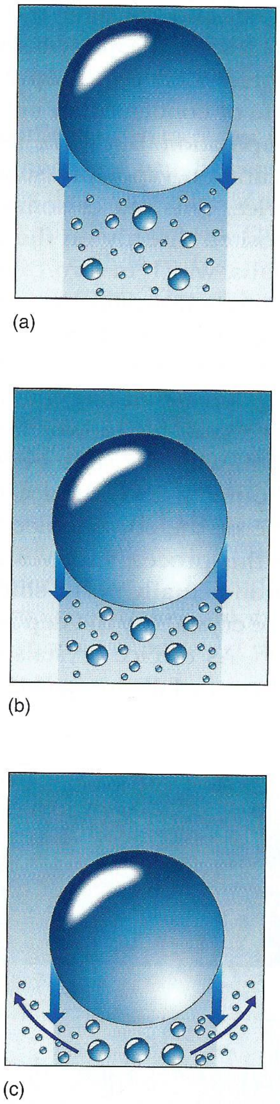
You have probably witnessed a similar phenomenon on a larger scale while driving down a country road in summer, with your windshield turning relatively large flying insects into "bug juice." Too heavy to get swept aside by the compressed air immediately ahead of the windshield, the bugs follow their own paths until the fateful moment of impact. Smaller bugs, in contrast, get blown out of harm's way.

Collision efficiencies are low for droplets nearly equal in size to the collector drop because their terminal velocities are so close to the collector's velocity that it is difficult for the collector to catch up to and collide with them. Continuing with the car analogy, collisions between vehicles are unlikely as long as all move at the same speed and direction.

Under certain situations, collision efficiencies can actually exceed 100 percent, and the collector can collide with more droplets than are in its path. A falling drop creates turbulence that can entrain small droplets outside its path and carry them back toward the top of the collector, where collision occurs.



▲ **FIGURE 7-4** Because of their greater mass, collector drops have greater terminal velocities (indicated by the length of the downward-pointing arrows) than do the smaller droplets in their path. Collector drops overtake and collide with the smaller ones.



▲ **FIGURE 7-5** As a collector drop falls (a), it compresses the air beneath it (b). This causes a pressure gradient to develop that pushes very small droplets out of its path (c). The small droplets get swept aside and avoid impact.

Recent research using mathematical models shows that turbulence in the form of whirling vortices greatly enhances collision efficiency. The vortices function like small centrifuges, separating droplets according to size as they spin around the center. The resulting variations in concentration significantly increase the average collision rate. In addition, rapid spinning causes jets of droplets to detach from the air flow like rocks thrown from a sling. The ejected droplets have a high probability of colliding with other droplets, so this process enhances collision efficiency. Calculations show that only mild turbulence is required for the centrifuge and sling effects, which implies that the processes operate in most clouds.

Coalescence When a collector drop and a smaller drop collide, they can either combine to form a single, larger droplet or bounce apart. Most often the colliding droplets stick together. This process is called **coalescence**, and the percentage of colliding droplets that join together is the *coalescence efficiency*. Because most collisions result in coalescence, coalescence efficiencies are often assumed to be near 100 percent.

Collision and coalescence together form the primary mechanism for precipitation in the tropics, where warm clouds predominate. In the middle latitudes, most precipitating clouds have freezing temperatures, at least in their upper portions. This favors the growth of precipitation by another mechanism involving the coexistence of ice crystals and supercooled water droplets, the Bergeron process (also known as the Bergeron-Findeisen or ice crystal process) described in the next section.

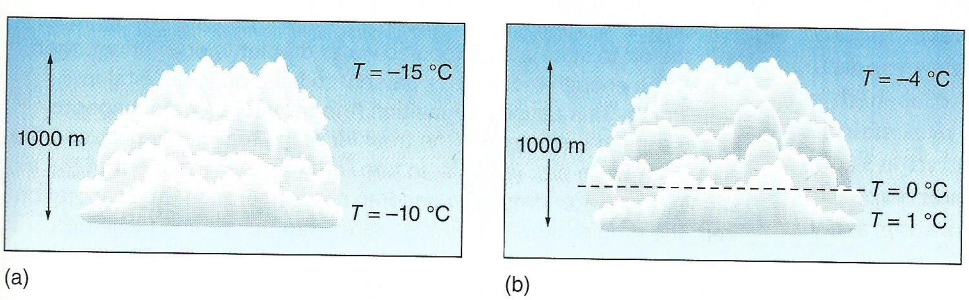
Checkpoint

1. **What are warm clouds and where are they most likely to be found?**
2. **What is the role of the collector drop in the formation of precipitation in a warm cloud?**

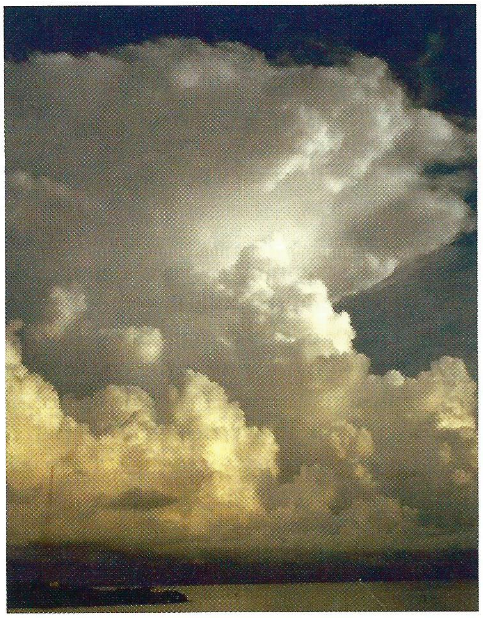
Growth in Cool and Cold Clouds

Unlike their counterparts in the tropics, at least a portion of most midlatitude clouds have temperatures below the melting point of ice. Some, such as the one in Figure 7–6a, have temperatures below 0 °C throughout and consist entirely of ice crystals, supercooled droplets, or a mixture of the two. These are referred to as **cold clouds**.

Cool clouds (Figure 7–6b), on the other hand, have temperatures above 0 °C in the lower reaches and subfreezing conditions above. As we discussed in Chapter 5, saturation at temperatures between about –4 °C (25 °F) and –40 °C (–40 °F) can lead to the formation of ice crystals, if ice nuclei are present, or to the formation of supercooled liquid droplets, if ice nuclei are absent. Thus, a well-developed cumulus cloud might be composed entirely of water droplets in its lower portion,



◀ **FIGURE 7-6** Cold clouds (a) have temperatures below 0 °C from their base to their top. Cool clouds (b) have temperatures above 0 °C in their lower portions with subfreezing temperatures above.



▲ **FIGURE 7–7** A cumulonimbus cloud. The lower portion consists entirely of liquid droplets, the middle a mixture of ice and liquid, and the upper portion entirely of ice. Note the less sharply defined margins of the glaciated portion composed of ice.

a combination of supercooled droplets and ice crystals in its middle section, and exclusively ice crystals in its upper reaches (Figure 7–7). The processes described in this section operate in cold and cool clouds having a mixture of ice and liquid water.

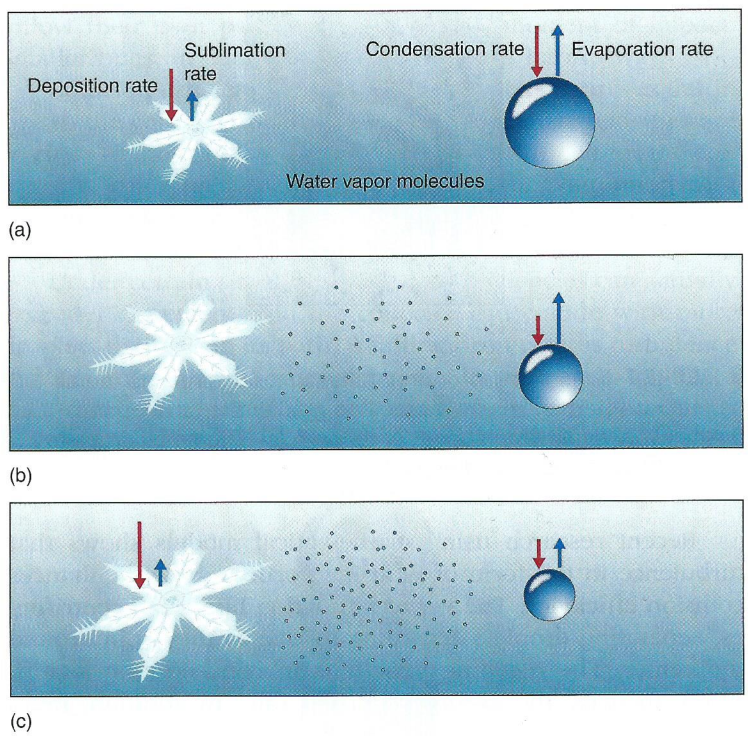
As we will now see, the coexistence of ice and supercooled water droplets is essential to the development of most precipitation outside the tropics. The process by which droplets and crystals in midlatitude clouds grow to precipitation size was first described by one of the preeminent figures of modern meteorology, Tor Bergeron. This process is therefore often referred to as the Bergeron process.

The principle underlying the **Bergeron process** is that the saturation vapor pressure over ice (the amount of water vapor needed to keep it in equilibrium) is less than that over supercooled water at the same temperature. This is because molecules in an ice crystal bond to each other more tightly than molecules of liquid water. You should recall that saturation exists when the vapor pressure of the air is at the point where evaporation from a water droplet would be exactly offset by condensation back onto it, or if sublimation from an ice crystal would be offset by deposition. Within a certain range of temperatures below zero Celsius, both ice crystals and supercooled droplets can coexist in a cloud. Ice crystals, however, do not sublimate ice to vapor as rapidly as water droplets evaporate liquid to vapor. Thus, ice crystals do not require as high a surrounding vapor pressure as do water droplets to remain in equilibrium, and they are said to have a lower saturation vapor pressure. As a result, if there is just enough water vapor in the air to keep a supercooled droplet from evaporating away, then there is more than enough water vapor to maintain an ice crystal. Let us see how that leads to precipitation.

Refer to Figure 7–8 and consider the situation in which ice crystals and supercooled droplets coexist, and the vapor pressure is equal to that needed to keep the droplets in equilibrium. In (a) the rate of condensation onto the liquid droplet equals the rate of evaporation. But while vapor pressure in the cloud equals the saturation vapor pressure for the droplet, it exceeds that for the ice. This causes some of the water vapor in the air to be deposited directly on the ice. The vapor content of the air then falls, which in turn causes the liquid droplet to evaporate as it gives up water to restore equilibrium (b).

The process does not end there, because evaporation from the droplet increases the water vapor content of the air, which causes further deposition onto the ice crystals (c). This leads to a continuous transfer in which the liquid droplets surrender water vapor, which is subsequently deposited onto the ice crystals. In other words, the ice crystals continually grow at the expense of the supercooled droplets. Although Figure 7–8 suggests this process involves distinct steps, evaporation and deposition actually occur simultaneously.

The growth of ice crystals by the deposition of water vapor initiates precipitation. As the ice crystals grow, their increasing mass enables them to fall through the cloud and collide



▲ **FIGURE 7–8** The Bergeron process. If exactly enough water vapor is in the air to keep a supercooled water droplet in equilibrium, then more than enough moisture is present to keep an ice crystal in equilibrium. This causes deposition (the transfer of water vapor to ice) to exceed sublimation (the transfer of ice to water vapor), and the crystal grows in size (a). This, in turn, draws water vapor out of the air, causing the water droplet to undergo net evaporation (b). Evaporation from the droplet puts more water vapor into the air and facilitates further growth of the ice crystal (c). Although this is shown here as a sequence of discrete steps, the processes occur simultaneously.

with droplets and other ice crystals. The collisions cause two other important processes to occur that greatly accelerate the growth rate of the ice crystals: riming and aggregation.

Riming and Aggregation We have seen that the formation of ice crystals in the atmosphere usually requires the presence of ice nuclei, or particles that initiate freezing. It so happens that ice itself is a very effective ice nucleus. Thus, when ice crystals fall through a cloud and collide with supercooled droplets, the liquid water freezes onto them. This process, called **riming** (or accretion), causes rapid growth of the ice crystals, which further increases their fall speeds and promotes further riming.

Another process in the development of precipitation is **aggregation**, the joining of two ice crystals to form a single, larger one. Aggregation occurs most easily when the ice crystals have a thin coating of liquid water to make them more "adhesive." Water is more likely to be present when the cloud temperature is not much below 0 °C, so adhesion is more common at the warmer end of cold clouds. (Perhaps you have noticed that very large snowflakes are more common during warm, early season snows, as opposed to those that come in the dead of winter.)

The combination of riming and aggregation allows ice crystals to grow much faster than by the deposition of water vapor to ice alone. In fact, growth rates from the three processes combined allow the formation of precipitation-sized crystals within about half an hour from the initial formation of the ice. When the ice crystals begin to fall, precipitation begins. What happens to these crystals as they fall determines the type of precipitation that occurs.

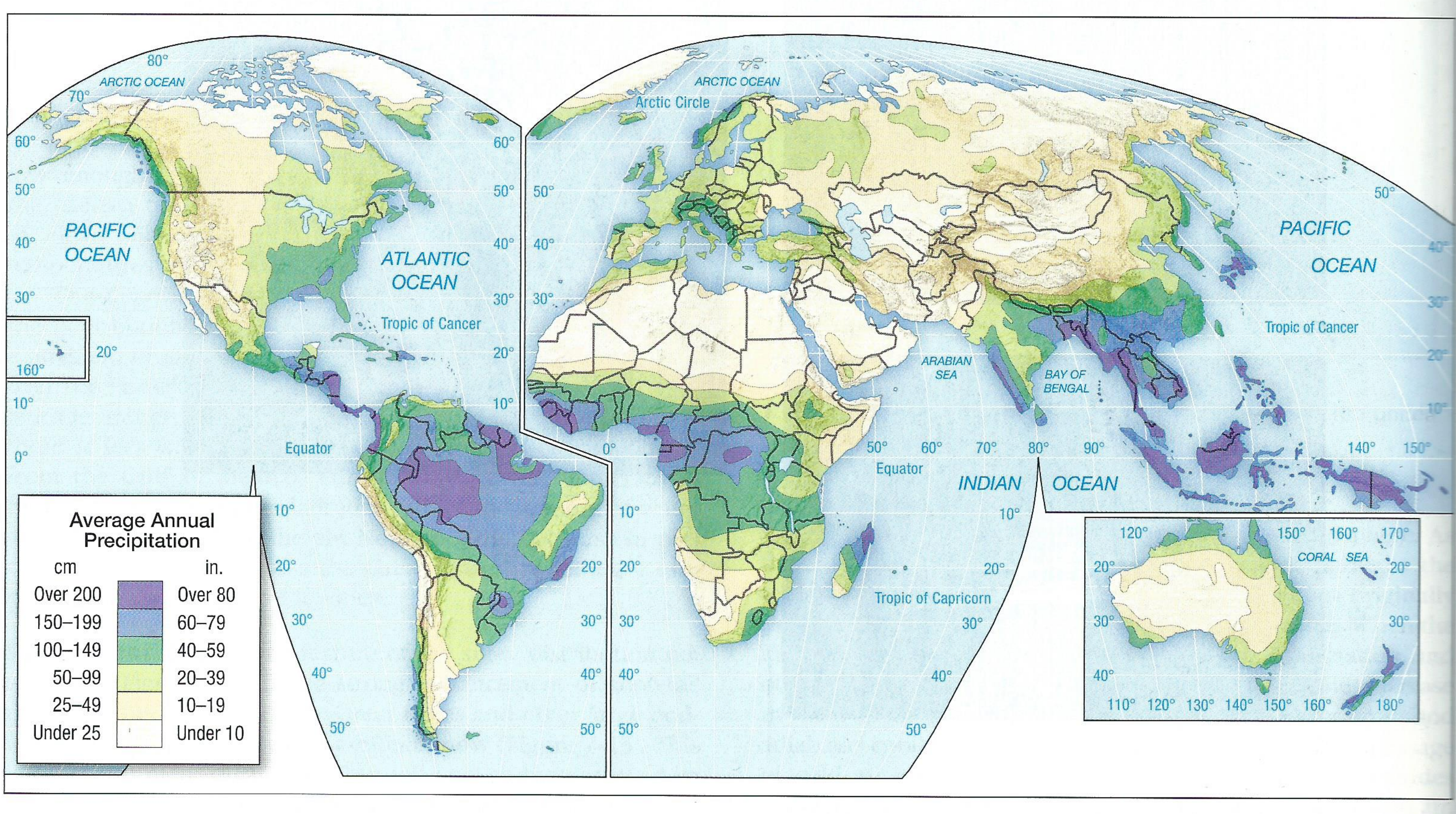
Checkpoint

1. What is the difference between a cool cloud and a cold cloud?
2. How does the fact that the saturation vapor pressure over ice is less than the saturation vapor pressure over supercooled water lead to precipitation?

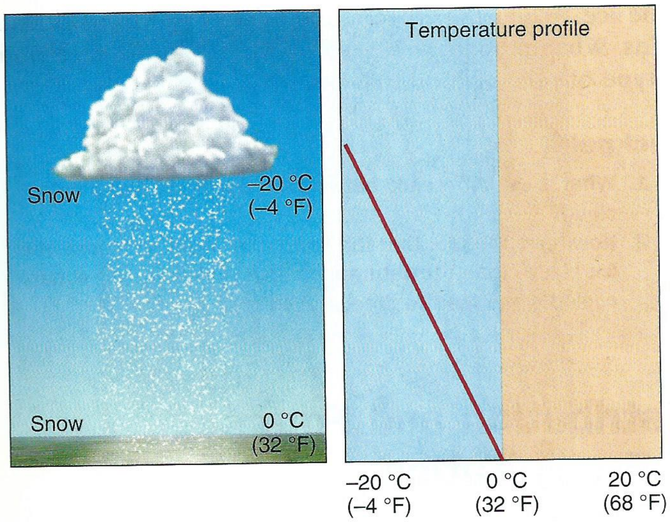
Distribution and Forms of Precipitation

The processes described above are ultimately responsible for all the various kinds of precipitation. Figure 7–9 shows the distribution of mean annual precipitation based on worldwide weather records.

In the tropics, precipitation occurs primarily by the collision–coalescence process, and it can therefore occur only as rain. In the middle latitudes, where ice crystal processes dominate, precipitation occurs as a solid or a liquid, depending on the temperature profile of the air through which it falls. If precipitation reaches the surface without ever having melted, we recognize it as snow. If it melts on the way down,



▲ **FIGURE 7–9** Average annual world precipitation.



▲ **FIGURE 7-10** Snow is initiated by the Bergeron process and reaches the ground only if the crystals falling to the ground never encounter temperatures above zero degrees Celsius.

it might reach the surface as rain. But raindrops sometimes freeze again before, or immediately after, reaching the surface, and then a different type of precipitation results. We now discuss the various types of precipitation.

TUTORIAL
PRECIPITATION

Use the tutorial to study saturation vapor pressure over ice and observe all three growth processes in cold clouds.

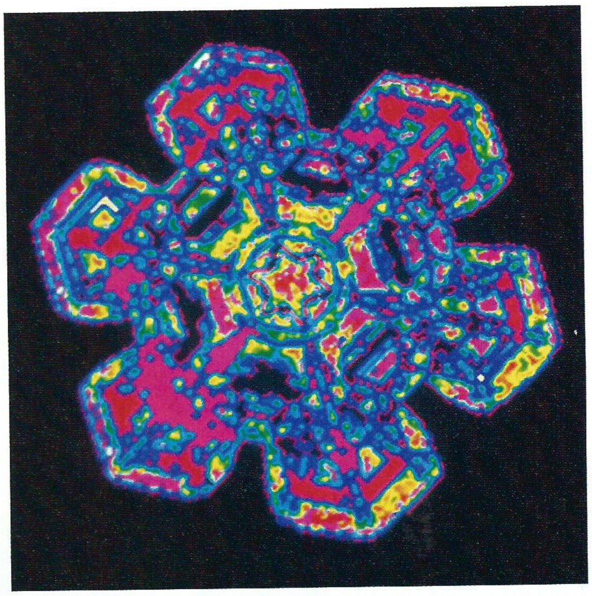
Snow

Precipitation as **snow** is initiated by the Bergeron process and the growth of crystals by riming and aggregation. Snow occurs if the falling crystals do not melt prior to reaching the ground. Therefore, cloud temperatures must be less than −4 °C (25 °F) and temperatures from the surface to the cloud base not much more than 0 °C (snow might not completely melt if it falls through a shallow layer of air not much warmer than 0 °C). Figure 7–10 illustrates a typical temperature profile required for snowfall.

Ice crystals in clouds can have a wide variety of shapes, including six-sided plates, columns, solid or hollow needles, and complex dendrites with numerous long, narrow extensions (Figure 7–11). The structure depends on the temperature and moisture conditions that exist when the crystal is formed.

If all of a crystal's growth occurs under similar conditions, its structure can be quite simple. If, on the other hand, the temperature and moisture conditions change during growth, a complex mixture of plate, needle, and dendrite can

(a)



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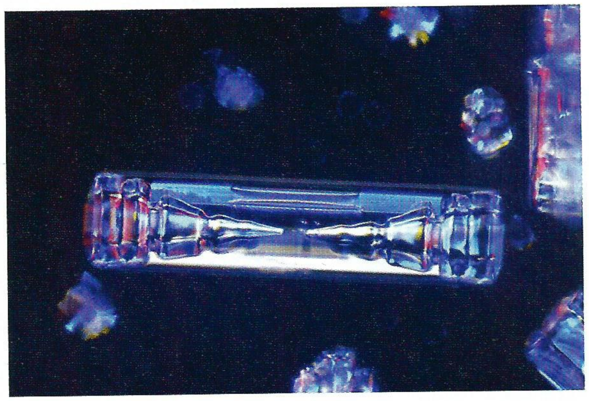


(c)

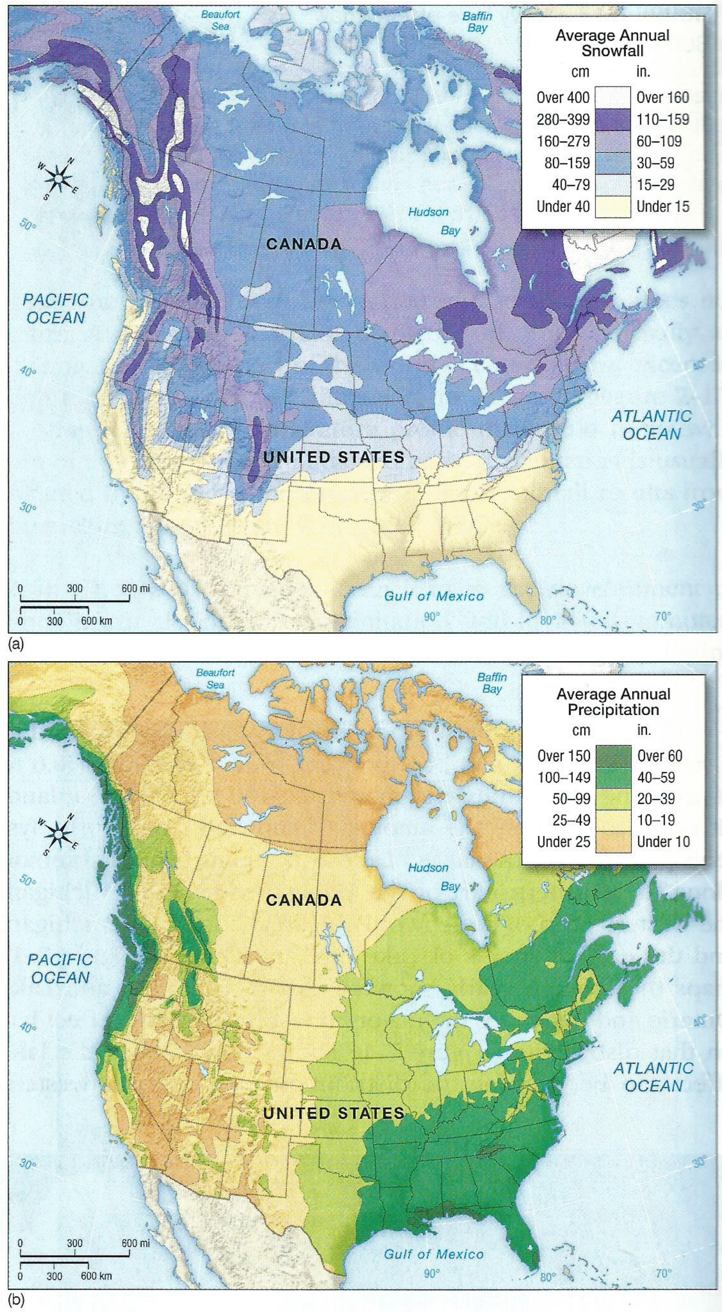
▲ **FIGURE 7-11** Ice crystals can assume several general shapes, including dendrites (a), plates (b), and columns (c). Each is favored under certain conditions of moisture content and temperature.

develop. Consider, for example, a crystal that originates in the cold, upper reaches of a cloud and gradually falls through a warmer environment. Because each combination of moisture and temperature tends to favor a different type of structure, the crystal can have a particular form at its nucleus, with other forms superimposed.

Snowflakes exist in a wide range of shapes and sizes. They can be as small as about 50 μm or as large as 5 mm. Where riming is the dominant growth process (which is the case in relatively warm clouds), the crystals tend to form a dense "wet" snowpack, ideal for snowball fights but no friend to snowblowers. In contrast, very cold snow typically forms small snowflakes that accumulate on the ground with a lower density. Because of their low temperature, these crystals have less adhesion and are difficult to pack. Skiers know this type of snow as *powder*. There is a widely held misconception about snow. Some people believe it can be too cold to snow, but this is not the case. It can be too cold to snow *a lot*, but it is never too cold to snow *at all*. Because mass is conserved, any ice crystals that form must do so at the expense of the water vapor content of the air. At very low temperatures, only a small amount of water vapor can exist in the air. And without an ample supply of water vapor, cooling of the air can cause the deposition of only a limited supply of ice. It is still possible for some snow to occur, no matter how low the temperature.

North American Distribution Figure 7–12a maps the distribution of mean annual snowfall across Canada and the United States. In the western portion of North America, the distribution of snowfall is governed largely by the presence of north–south mountain ranges (the Coast Ranges, the Sierra Nevada, the Cascades, and the Rockies) that provide orographic uplift and enhance the precipitation from passing storm systems. At high elevations, these ranges have winter temperatures low enough that most precipitation occurs as snow. Over the eastern two-thirds of the continent, there is an increase in mean snowfall with latitude, mostly because the lower temperatures at higher latitudes favor snow rather than rain.

The distribution of annual snow is in marked contrast to the distribution of annual precipitation—rain plus the water equivalent of snow (Figure 7–12b). Total annual precipitation over the eastern two-thirds of North America decreases with latitude rather than increases, largely due to the fact that there is less water vapor in the air with increasing distance from the Gulf of Mexico. Furthermore, lower temperatures typically found at higher latitudes reduce the amount of water vapor that can exist in the air. Note, too, the decrease in precipitation westward across the Great Plains, revealing a rain shadow in the lee of the Rockies.

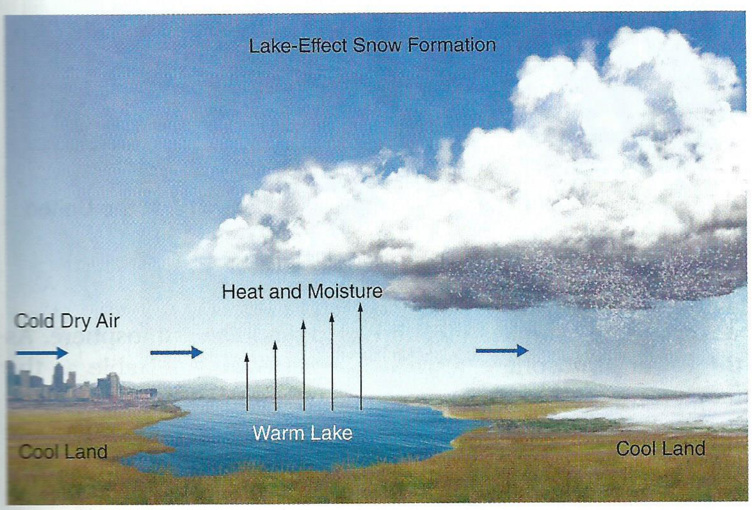


▲ **FIGURE 7–12** Average annual snowfall in Canada and the United States (a), and average annual precipitation (b).

Lake-Effect Snow One feature of the snow distribution not shown in Figure 7–12 is the strong enhancement of snowfall that occurs downwind of the Great Lakes and other large bodies of water, referred to as **lake-effect snow** (Figure 7–13). This is most prevalent during the late fall and early winter, when lake temperatures are still moderately high but cold air can pass over from the north. As shown in Figure 7–14, the lake warms and evaporates moisture into the lower atmosphere. As the lower atmosphere warms, it can become unstable as the temperature lapse rate increases. Thus, air that was originally dry and stable becomes moist and statically unstable. When the air passes over land, the effects of topography, vegetation, and other features of the land surface slow the wind. The decrease in wind speed causes convergence, a mechanism for uplift and adiabatic cooling discussed in Chapter 6. Thus, the passage of cold air over the lakes and subsequent landfall provides the three mechanisms favorable for precipitation: unstable air, sufficient water vapor, and a mechanism for uplift.

▸ **FIGURE 7–13** Heavy lake-effect snow in Buffalo, New York.

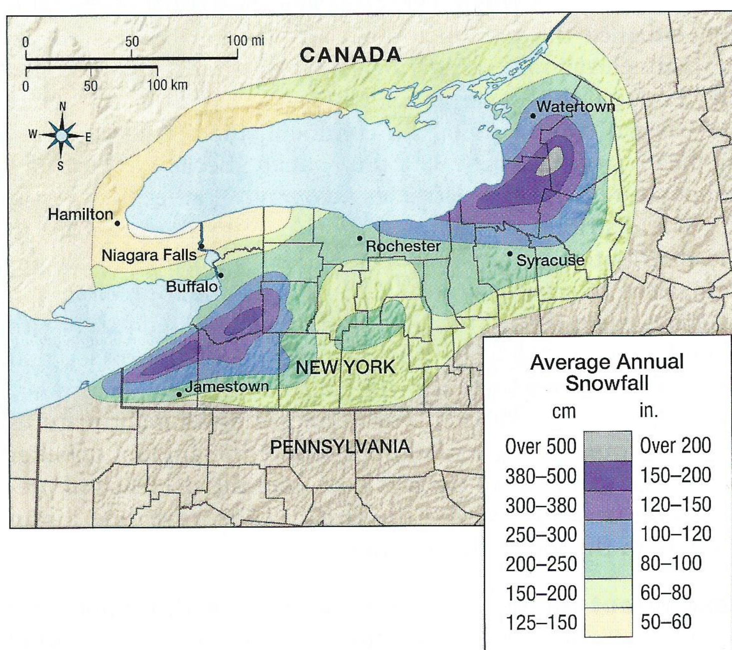
This lake effect often produces snow showers restricted to a strip of land that can be anywhere from 1 to 80 km (0.6 to 50 mi) long and can extend more than 100 km (60 mi) inland. (It can also increase the amount of snowfall from storm systems passing over the lakes.) Lake-effect snow is most common along the northern part of the Upper Peninsula of Michigan, the western strip of the Lower Peninsula along Lake Michigan, and the eastern shores of Lake Erie and Ontario. Figure 7–15 maps the average yearly snowfall east of Lake Erie and Lake Ontario and illustrates the major impact that the lake effect has on that distribution. Figure 7–16 shows how intense the lake effect can be, plotting the distribution of snow over western



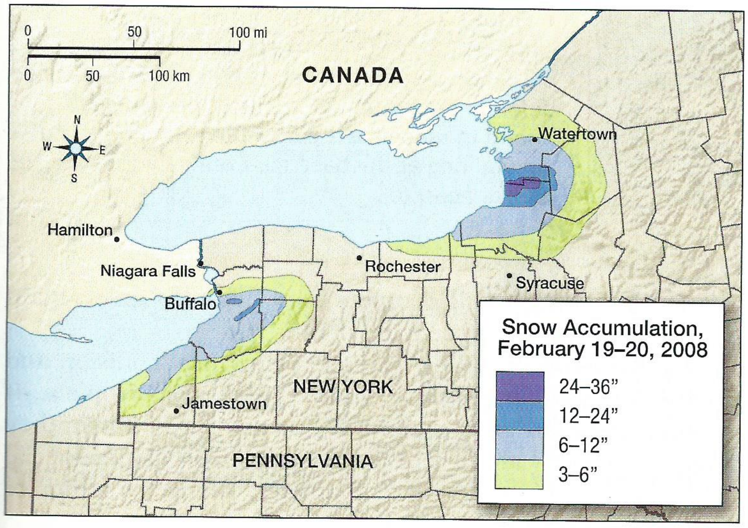
▲ **FIGURE 7–14** Lake-effect snow formation. As cold air moves over a warm lake, heat and moisture are brought into its base. The moist, unstable air mass is then subject to surface convergence as it slows down upon reaching the downwind shoreline, resulting in sometimes heavy snowfalls.

Did You Know?

The greatest recorded seasonal snowfall occurred at Mt. Baker Lodge, Washington, at 1500 m elevation (5000 ft) over the winter of 1998–99. The total of 2736 cm (1140 in., or 90 ft!) exceeded the previous record observed at Mt. Rainier, Washington, during the winter of 1971–72 (2693 cm, or 1122 in.).



▲ **FIGURE 7–15** The average seasonal distribution of snow east of Lake Erie and Lake Ontario (inches). Note the increase in snow cover immediately downwind of the lakes.



▲ **FIGURE 7–16** Heavy lake-effect snow in February 2008. Snowfall locally exceeded 1 m (3.3 ft) in depth, while much of the general area received no snow.

and upstate New York over the 2-day period, February 19–20, 2008. Notice how large amounts of snow fell over two limited areas while nearby regions got no snow at all.

The winter of 1976–77 provided one of the most noteworthy seasons for lake effect snowfall in the Great Lakes region, when 51 days of lake-effect snow produced record accumulations in upstate New York. The 103 cm (40.5 in.) of snow that fell in a 4-day period from late November to early December in Buffalo provided only a preview of what was to come. By the end of January, Buffalo had received 3.6 m (12.5 ft) of snow for the 3-month period beginning November 1. Even greater accumulations occurred in northern New York, downwind of Lake Ontario, where up to 9.5 m (33 ft) of snow fell over the course of the winter.

Lake-effect snowfall can also occur on the northern side of Lake Ontario. In January 1999, for instance, Toronto experienced a series of storms that brought almost 120 cm (50 in.) of snowfall—an all-time monthly record for the city. During the first 2 weeks of the month, a normal year's worth of snow fell.

Checkpoint

1. What is lake-effect snow and where does it typically occur in the United States?
2. How does the movement of cold, dry air over a warm lake produce lake-effect snow?

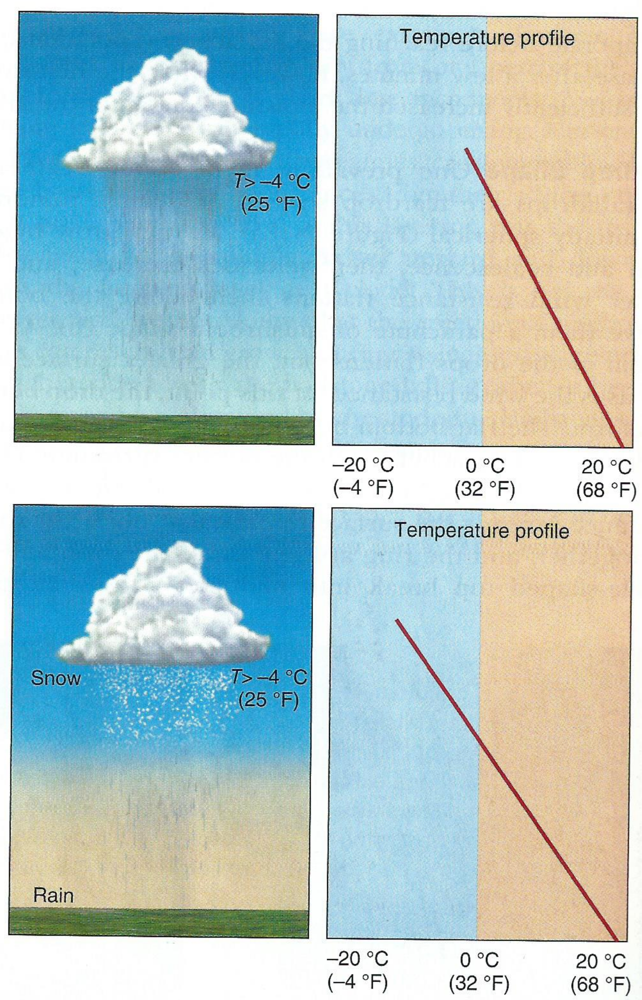
Rain

As we have already seen, most precipitation in the tropics comes from warm clouds whose temperatures are somewhat above the melting point of ice. Furthermore, the air temperature below the clouds is well above freezing, so the rain does not freeze after leaving the base of the cloud. Thus, virtually all precipitation in this part of the world occurs as **rain**, except in high mountains such as Kilimanjaro in Tanzania. Figure 7–17a illustrates a typical temperature profile that could occur with rain of this type. In the middle latitudes, precipitation is usually initiated by the Bergeron process, so most rainfall results from the melting of falling snow (Figure 7–17b).

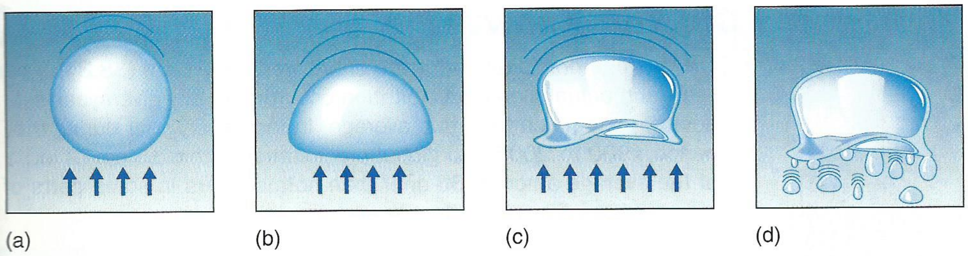
Did You Know?

The average precipitation for the entire world is 97 cm (38.8 in.) per year, and actual global precipitation occurring in any given year seldom departs much from this average. In fact, every year during the period 1900 to 2005 had global precipitation within ±5 cm (2 in.) of the average amount. So drier than normal years in some parts of the world are generally offset by wet conditions elsewhere.

Rain Showers Convection can lead to the development of cumuliform clouds and precipitation within a few minutes. The episodic precipitation from these rapidly developing



▲ **FIGURE 7–17** Rain can fall when the collision–coalescence process occurs (a), or when falling ice crystals fall from a cloud and melt on the way to the ground (b).



▶ **FIGURE 7–18** Raindrops are not teardrop shaped. They are initially spherical (a) but flatten out on the bottom as they fall (b). Wind resistance deforms the bottom of the drop, stretching it inward to resemble a parachute with a doughnut-shaped ring at the base (c). Eventually, the droplet breaks apart (d).

clouds is called **showers**. Showers can occur as either rain or snow, but because convection is usually most vigorous in the warm season, showers are more likely to occur as rain.

During a steady rain, droplets occur in a wide variety of sizes. In a shower, the first droplets all tend to be large and widely spaced, but within a short period of time the large droplets give way to a greater number of smaller ones. What happens is quite simple: Large and small droplets fall from the base of the cloud together, but the larger raindrops have greater terminal velocities and reach the surface while the smaller ones are still falling through the air.

Another factor favors the occurrence of large drops at the beginning of a rain shower. Because they take longer to fall through the unsaturated air, small raindrops are more likely to evaporate before reaching the surface. (Evaporation does decrease after a few minutes, however, when the first drops have sufficiently increased the moisture content of the air.)

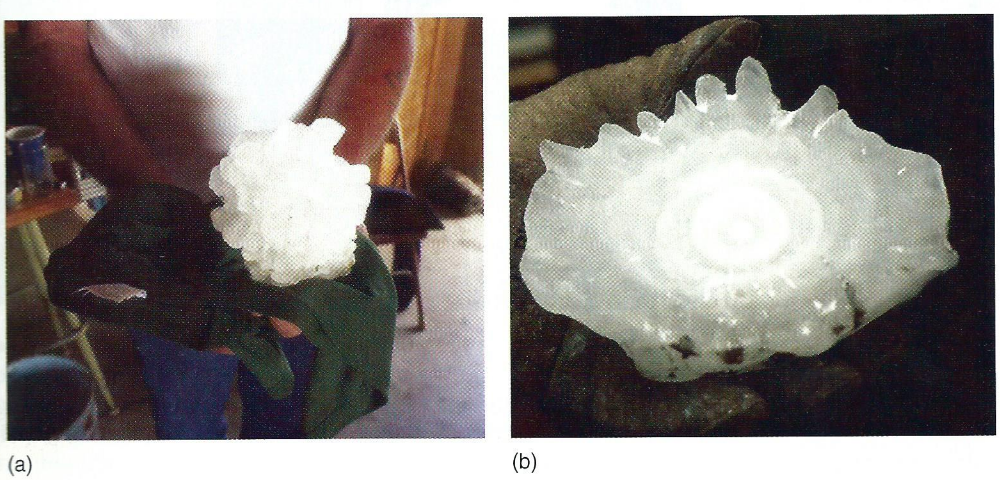
Raindrop Shape One prevailing myth about weather is that raindrops are teardrop shaped. In reality, raindrops are initially spherical (Figure 7–18a). As they grow by collision and coalescence, their velocities increase, and the greater wind resistance flattens them along the bottom to give them a parachute or mushroom shape (b). As the bottom of the drops flattens out, the greater surface area increases the wind resistance. At this point, the drop begins to deform. The flat bottom becomes concave, somewhat in the shape of a parachute with the bottom surrounded by a relatively thick, doughnut-shaped rim (c). Eventually, wind resistance exceeds the surface tension that holds the droplets together, and the ring at the base of the drop and the bubble-shaped top break into multiple, smaller droplets (d). The resulting small droplets then begin to grow again by collision and coalescence.

The breakup of falling drops explains why collision and coalescence do not produce enormously large droplets. If the drops were to grow continually on their way down, they could conceivably attain the size of basketballs! Under special conditions, droplets can have diameters of up to 5 mm or more, but they are seldom larger.

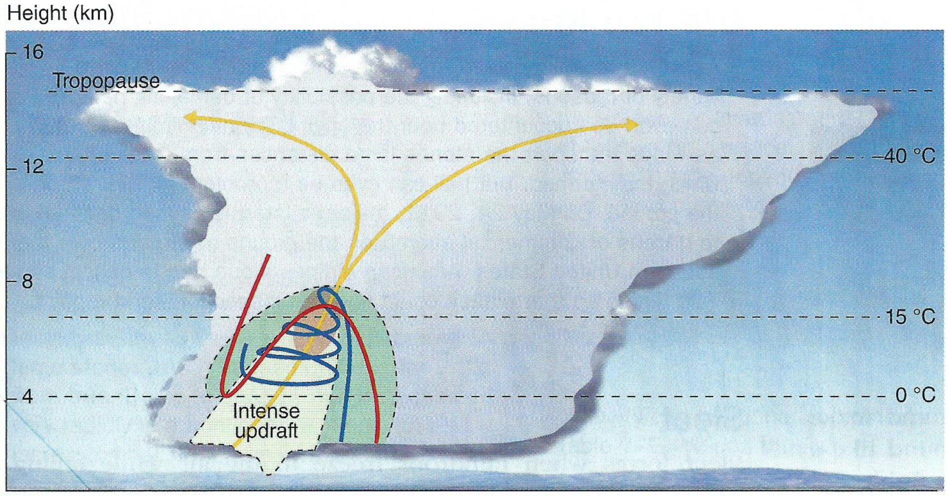
Graupel and Hail

When an ice crystal takes on additional mass by riming, its original six-sided structure becomes obscured and its sharp edges are smoothed out. The new ice may contain very small air bubbles that give it a spongy texture and milky-white appearance. This type of modified ice crystal is called **graupel** (pronounced GRAU-pull). Graupel pellets attain diameters up to 5 mm or so, giving them terminal velocities of about 2.5 m/sec (5 mph). Graupel pellets can fall to the ground as precipitation, but under other circumstances they can remain in the cloud and provide the nuclei upon which hailstones form.

Hail consists of multiple layers of ice that are usually no more than a few millimeters in thickness, but under extreme circumstances can achieve sizes comparable to softballs (Figure 7–19). Hailstones are often relatively concentric, though some have irregular shapes. No other type of precipitation is capable of producing drops or particles nearly as large as hail, so it should not be surprising that large amounts of water must be available in the cloud in which they grow and that very strong updrafts are required to allow precipitation this large to remain airborne. Hail therefore forms from cumulonimbus clouds.



▶ **FIGURE 7–19** Though usually no larger than pea-sized, hail can be remarkably large (a). Its growth accrues by the repeated addition of new layers of ice (b).



◄ **FIGURE 7–20** The formation of hail. Embryos near zones of strong updrafts may be blown into the updraft, rapidly lifted upward, and ejected as small hail. Hail that forms as it rotates around these updrafts can accumulate more ice to become large and fall through a region called a hail cascade.

Hail development begins with the formation of a small particle called an *embryo*, usually consisting of graupel, in a region of cloud where supercooled droplets coexist with ice crystals (Figure 7–20). The cumulonimbus clouds that produce hail do not have uniformly strong updrafts across their entire width; they have localized areas of extremely violent vertical motions surrounded by zones of relatively weak updrafts. Hail growth typically begins just outside these intense updrafts.

Some developing hailstones get blown from the edge of the strong updrafts into the core, where they are rapidly taken to the upper reaches of the cloud and ejected. When this happens the stones do not have time to accumulate large amounts of ice and fall from the cloud as relatively small hail.

Large hail can develop if the growing particles rise slowly around the more intense core of the updraft. This provides time for the hail to come into contact with supercooled liquid droplets that freeze onto the existing ice. This process can take place in one of two ways: the *dry growth* or *wet growth regimes*.

The dry growth regime occurs at temperatures well below 0 °C (32 °F), wherein supercooled water droplets collide with and freeze onto ice crystals. As the water freezes it releases latent heat that warms the crystal, but not sufficiently to bring its temperature to 0 °C.[1] The ice forms rapidly, allowing it to incorporate very small air bubbles that give it a whitish, opaque appearance.

The wet growth regime from the latent heat released when water freezes onto the ice is sufficient to bring its temperature to 0 °C. The water can remain as a liquid for some time, during which the air bubbles can be released. When the water does freeze, the lack of air bubbles gives it a much clearer, translucent appearance. The existence of supercooled water also allows the ambient wind conditions to deform the layer of water so that when it freezes it takes on a nonconcentric shape, as in Figure 7–19a.

Cumulonimbus clouds have extremely dynamic conditions, and things change readily from one part of the cloud to another and over very short time increments. As a result, growing hailstones can rapidly undergo changes in environment between the wet and dry regimes, causing multiple layers of ice having different appearances from their adjacent ones.

It was once believed that this layering was the result of hailstones making multiple passes upward and downward across the freezing level in a cloud. This is not the case, however, and we now know that the creation of multiple ice layers can result from growth during a single passage upward through a cloud, with the hailstone falling when it becomes too heavy to be maintained by the updrafts. Hailstones often fall in a localized area outside the zone of strongest updrafts, called a **hail cascade**.

Although most hailstones are pea sized, they can become as large as marbles, golf balls, or, under the most violent of conditions, even softballs!

Did You Know?

The largest hailstone ever recorded was found in Vivian, South Dakota, on July 23, 2010. It was officially measured at 19 cm (8 in.) in diameter and weighed 0.88 kg (1 pound, 15 ounces). A local ranch hand found the hailstone and immediately stored it in a freezer. Apparently the stone was even larger when initially recovered, but the storm knocked out power and the stone partially melted before it could be measured.

Hailstones consist mainly of ice with only a small amount of air mixed in. Because ice is relatively dense—90 percent as dense as liquid water—hailstones can become fairly heavy. Compare hailstones to snowflakes, whose volume is occupied mostly by air. Snowflakes have relatively little mass and low

[1]This latent heat of fusion is the heat released when liquid water freezes. It is analogous to the latent heat of evaporation released when water vapor condenses to form a liquid.

TABLE 7–1

Terminal Velocities of Hailstones

$V_t = \sqrt{\frac{1}{3}\rho/k}\,\sqrt{r} = 20\sqrt{r}$ (r in cm, V_t in m/sec)

Radius (cm)	Terminal Velocity (Meters per Second)
0.1	6 (13 mph)
1.0	20 (44 mph)
2.0	28 (62 mph)
3.0	35 (77 mph)

Did You Know?

Aviators are well advised to avoid cumulonimbus clouds for a variety of reasons, including the possibility of damaging hail. Hail can even be encountered near the anvil of cumulonimbus clouds as winds transport the stones large distances from the area in which they formed. But hail can even be a problem for aircraft on the ground. On May 24, 2011, major hailstorms caused damage to dozens of commercial aircraft on the ground at airports over the southern United States. American Airlines alone had to cancel some 700 flights so that planes could undergo inspections and repairs.

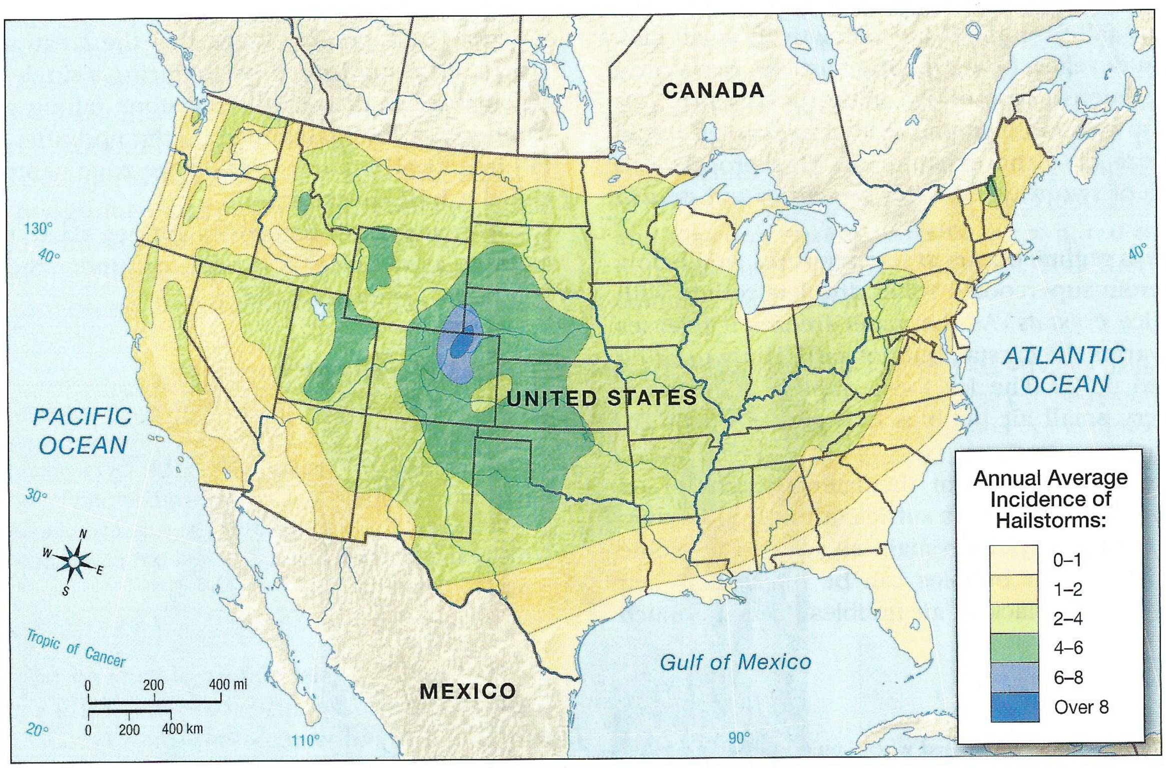
terminal velocities, so they flutter to the ground and make barely a sound. Hailstones, on the other hand, sound like a barrage of falling marbles as they hit the surface. Table 7–1 lists the terminal velocities of various sizes of hailstones.

Hailstones the size of baseballs (radius = 3.5 cm) contain about 160 g of ice (weighing about a third of a pound) and fall at about 40 m/sec (88 mph)! No wonder they are capable of producing tremendous damage. Hailstorms present a major threat to the Great Plains of the United States (Figure 7–21) and Canada (Alberta and interior British Columbia), where they are known to destroy entire fields of crops in a matter of minutes. Because they are most common in the spring and summer, it is often too late to replant the acreage with new seed. (See *Box 7–2, Physical Principles: The Effect of Hail Size on Damage,* for more information about the destructive potential of hail.)

Sleet

Sleet forms when raindrops freeze in the air while falling. Because most rain outside the tropics originates from the ice crystal process, sleet begins as falling ice crystals or snowflakes. As the ice falls through the air, it encounters warmer air and melts to form a raindrop. If the falling raindrop then encounters a lower layer of air whose temperature is below 0 °C, it can refreeze to form sleet. This process, shown in Figure 7–22, results in semitransparent pellets smaller than about 5 mm (0.2 in.) in diameter. Because the formation of sleet requires that a droplet fall through air that is cooler near the surface than aloft, it necessarily requires an inversion, usually one associated with a warm front (which we describe in Chapter 9).

Of course, a raindrop will not freeze instantaneously; sufficient cooling must take place as it falls through the



▲ **FIGURE 7–21** The annual average number of hailstorms over the United States.

7–2 FOCUS ON SEVERE WEATHER

The Effect of Hail Size on Damage

You might wonder why large hailstones are more damaging than small ones. After all, isn't the issue how much ice falls to the ground, and aren't many small stones roughly equivalent to a smaller number of large stones? As it happens, this is far from true. Damage done by hail increases very rapidly in a nonlinear fashion with increasing hailstone size. What matters is the amount of kinetic energy hail carries as it falls to the surface.

Kinetic energy depends on mass (m) and speed (v) according to

$$KE = \frac{1}{2}mv^2$$

The mass of a stone is its density (ρ) times its volume

$$\left(\frac{4}{3}\rho\pi r^3\right)$$

Velocity can be found using the formula in Table 7–1. We can thus write

$$KE = \frac{1}{2}\left(\rho\frac{4}{3}\pi r^3\right)(20^2 r)$$

We see that the kinetic energy of a hailstone is proportional to the fourth power of its radius. As a result, a hailstone with a radius of 1 cm (0.4 in.) packs not twice as much punch as a 0.5 cm (0.2 in.) hailstone, but 16 times as much. The threshold for automobile windshield damage due to falling hailstones is about 5 cm (2 in.) in diameter.

surrounding air. Thus, for sleet to develop, the layer of cold air beneath an inversion must be fairly deep. If it is too shallow, another type of frozen precipitation will occur, freezing rain.

Freezing Rain

Freezing rain (Figure 7–23) is one of the more deceptive weather events. It usually looks like a gentle rain—certainly nothing to cause major problems. In reality, widespread episodes of freezing rain (often referred to as **ice storms**) can literally paralyze transportation and communications for hundreds of square kilometers.

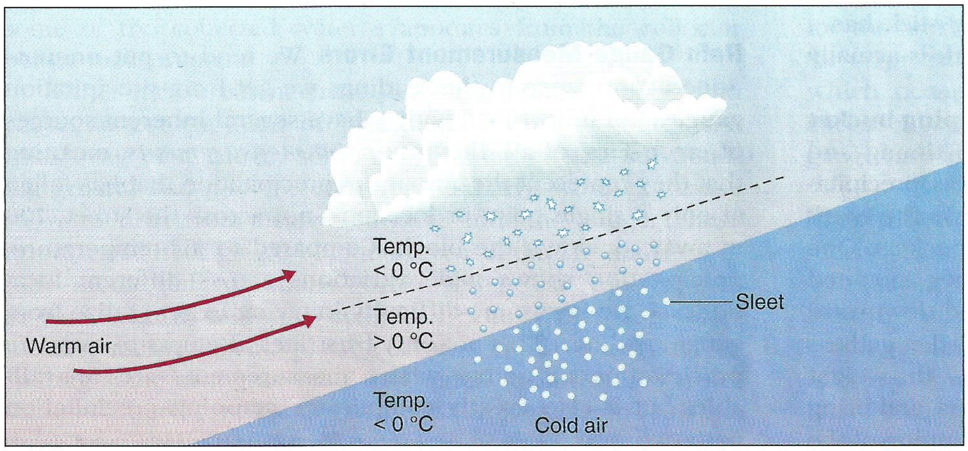
Freezing rain begins when a light rain or drizzle of supercooled drops falls through air with a temperature at or slightly below 0 °C. When the raindrops hit the surface, they form a thin film of water, but only for a moment. Soon afterward the water freezes to form a slick, continuous coating of ice.

When freezing rain hits roadways, the loss of friction that results leads to extremely dangerous conditions. The weight of accumulated ice also can cause tree limbs and telephone and power lines to snap and fall to the ground. When you imagine downed lines, impassable roads, and broken debris scattered about, it is easy to see why freezing rain can be so disruptive to human life. To make matters worse, freezing rain is often associated with slowly moving storm systems and may therefore persist for several days.

One of the most destructive ice storms in recent decades occurred from the Great Lakes through the southern Great Plains on December 9–11, 2007. Tens of thousands of homes and businesses lost power across Kansas, Missouri, Iowa, and Illinois, but the storm delivered its strongest punch to Oklahoma, where up to 3.5 cm (1.5 in.) of ice accumulated across the state. The entire state was declared a federal disaster area as 640,000 customers lost electrical power, some of them for up to a week. At least 27 people died from the storms, mostly from traffic accidents on the slick roads. If there were one favorable outcome from the event, it was the postponement of final exams across many of the colleges and universities in the Great Plains.

Checkpoint

1. Briefly define graupel, hail, sleet, and freezing rain.
2. How are the conditions under which sleet and freezing rain form similar? How are they different?



◄ **FIGURE 7–22** Sleet occurs as rain, falling from a cloud, passes through a cold layer and freezes into small pellets. This is most common along warm fronts.

▲ **FIGURE 7–23** Freezing rain.

Measuring Precipitation

Given the effects of precipitation on everyday activities, it is no surprise that we measure precipitation at many locations. Just as precipitation occurs in several forms, different types of gauges exist for measuring it. Each method has its own advantages and disadvantages.

Rain Gauges

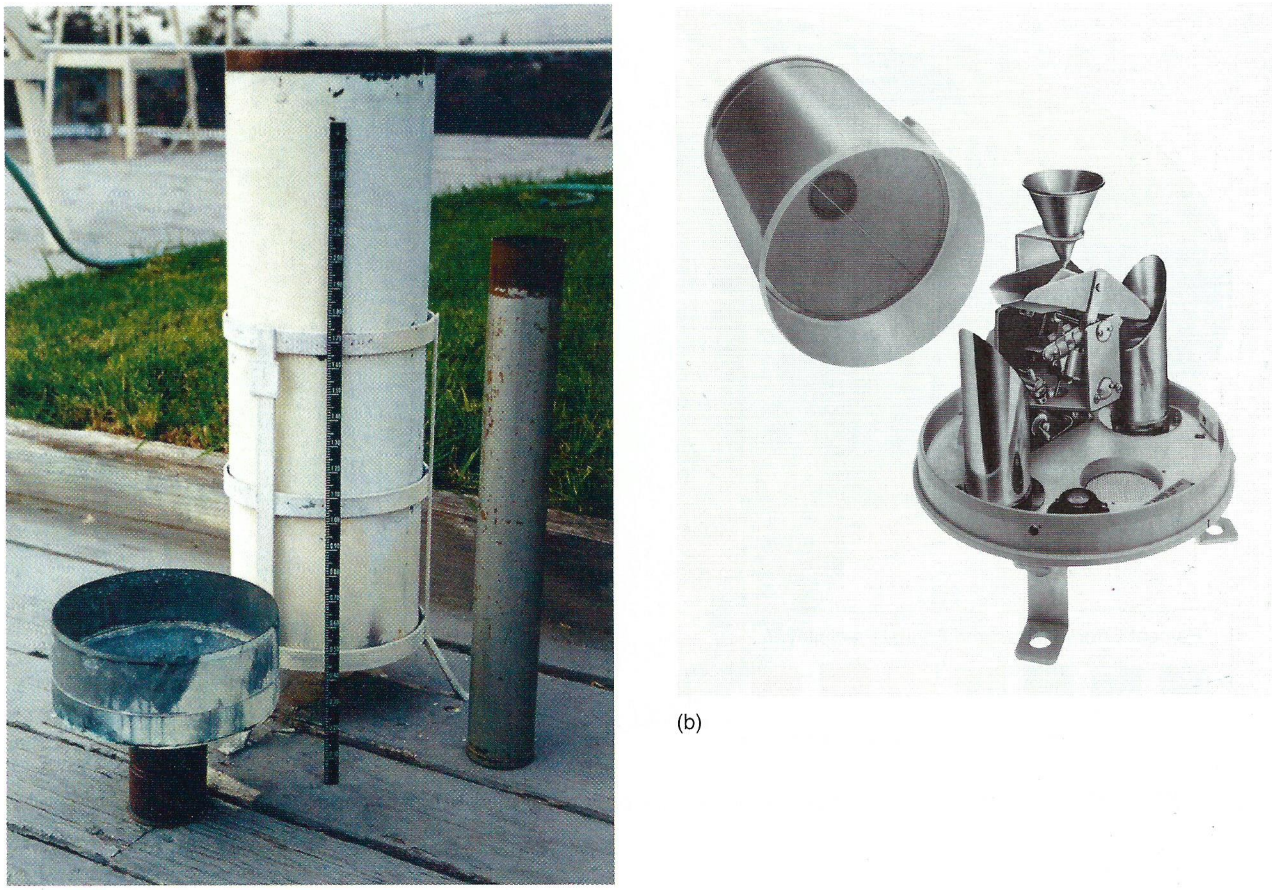
Rainfall is usually measured with a **rain gauge** (Figure 7–24a). Standard gauges have collecting surfaces with diameters of 20.3 cm (8 in.). The precipitation funnels into a tube with one-tenth the surface area of the collector, so the depth of accumulated water undergoes a tenfold increase. This amplification lets us measure the precipitation level precisely by simply inserting a calibrated stick into the water, removing it, and noting the depth of the wet portion, rather like checking oil with a dipstick. Note that the measuring stick has a correspondingly graduated scale so the 1 cm mark is actually 10 cm from the base of the scale.

An automated collector known as a **tipping-bucket gauge** (Figure 7–24b) provides a record of the timing and intensity of precipitation. This instrument funnels precipitation from the top like a standard rain gauge, but as the water accumulates it is stored in one of two pivoting buckets. One of the buckets is initially upright, while the other, mounted on the opposite end of a pivoting lever, is tipped downward and away from the collector. When the upright bucket gathers rain equivalent to a certain depth (usually 0.01 in.), the weight of the water causes it to tip over, empty its contents, and bring the opposite bucket to the upright position. The tipping of the pivoting buckets triggers an electrical current to a computer that precisely notes the time of the event. The number of tips per unit of time indicates the precipitation intensity. Older recorders use a rotating drum and a printer to provide an analog record of the precipitation rate.

Weighing-bucket rain gauges are similar to tipping-bucket gauges insofar as new accumulations of rain are constantly recorded. A weighing mechanism in these devices translates the weight of the accumulated water in the gauge to a precipitation depth, and the information is stored automatically.

Rain gauges are found at virtually all weather-recording stations, which makes precipitation data plentiful in economically developed countries. Data are scarce in much of the rest of the world, especially over the more than 70 percent of Earth's surface covered by ocean. Furthermore, measurement accuracy is a concern (even with modern instruments), due to such problems as evaporation from the gauge and winds that can prevent rain from entering the gauge.

Rain Gauge Measurement Errors We tend to put unquestioned confidence in the readings we get from precipitation gauges, but unfortunately they have several inherent sources of error. First of all, they are *point measurements*, meaning that they represent the amount of precipitation that has fallen at only a single point or location—not across the street, 100 m away, or down the block. Compared to air temperature, precipitation shows wide variations across different locations, so we face some difficulty in trying to generalize from gauge readings. This not only frustrates attempts to estimate precipitation at points where measurements are unavailable, but it also greatly complicates mapping precipitation patterns.



(a)

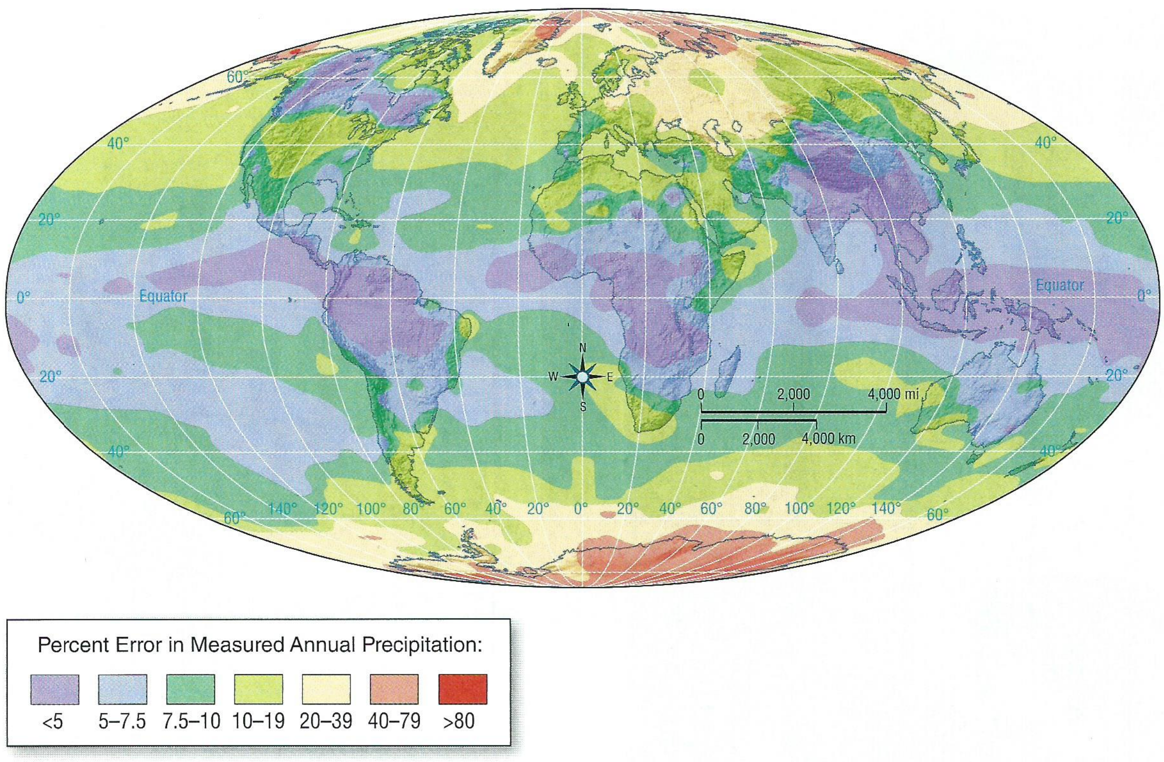
(b)

▲ **FIGURE 7–24** A standard rain gauge with its component parts (a). Rain captured by the collector (bottom left) is funneled into the narrow tube (right). The calibrating stick is inserted into the collection tube, and the length of the wetted portion indicates the precipitation accumulation. The interior workings of a tipping-bucket gauge are shown in (b).

Precipitation gauges have other flaws. Wind-generated turbulence near the top of a gauge deflects precipitation away from the collecting surface, leading to an underestimate of the true value (especially for snow). Of the precipitation that does find its way into the gauge, a certain amount goes unrecorded—some because the water splashes out on impact, and some because water is retained as a thin film that does not accumulate at the base of the collector. On hot, windy days some of the collected water evaporates from the collector before measurement.

Other measurement errors have the opposite effect and cause an overestimate of precipitation. Just as water can splash out of a gauge, some can bounce off the surrounding ground and into the collector. Similarly, wind can cause snow to drift from nearby surfaces into the gauge. These errors can affect gauge amounts even under the most carefully monitored conditions. Other things can cause errors as well, such as failing to completely empty the gauge after the last measurement, placing it on a nonlevel surface, or spilling some of the accumulated water. At well-run weather stations where conditions can be controlled, errors such as these are usually very small. Imagine, however, the difficulty in making reliable observations of precipitation on a ship or buoy. The collection surface of the gauge can be kept only as horizontal as a pitching and rolling ship deck will allow, and it is extremely difficult to prevent windblown seawater from entering the gauge. The moral here is that the distribution of precipitation is not as well recorded as we might suppose, nor as well recorded as befits its extreme value to agriculture and human welfare. Figure 7–25 shows estimated gauge errors for the world, based on a study that developed techniques for error estimation and correction. The most severe errors, which occur at high latitudes and over water, sometimes exceed 80 percent of the true values. Clearly, uncorrected gauge measurements must be used cautiously.

Precipitation Measurement by Weather Radar During recent years, the network of rain gauges has been augmented by radar measurements. Though radar will be discussed in more detail in Chapter 11, for now we can say that weather radars estimate the intensity of precipitation by emitting microwave radiation with wavelengths of several centimeters. Precipitating droplets, ice crystals, and hailstones scatter some of the emitted radiation back to the radar unit, which



▲ **FIGURE 7–25** The global distribution of the percent error in measured annual precipitation.

records the intensity of the backscattered radiation. In general, the more intense the backscattered radiation, the more intense the precipitation. Meteorologists have developed schemes that relate the intensity of backscattered radiation to the rate of precipitation.

Radiation is not emitted continuously by the transmitter, but just for very brief periods that are interspersed with momentary pauses. Sufficient time is allowed for each pulse of radiation to echo back to the transmitter/receiver unit before the next beam is emitted. The closer the precipitation is to the radar, the quicker the pulse will return to the unit. By measuring the strength of the return radiation and the time taken for it to return to the unit, a profile can be taken showing how much precipitation is occurring and how far from the radar it is.

Did You Know?

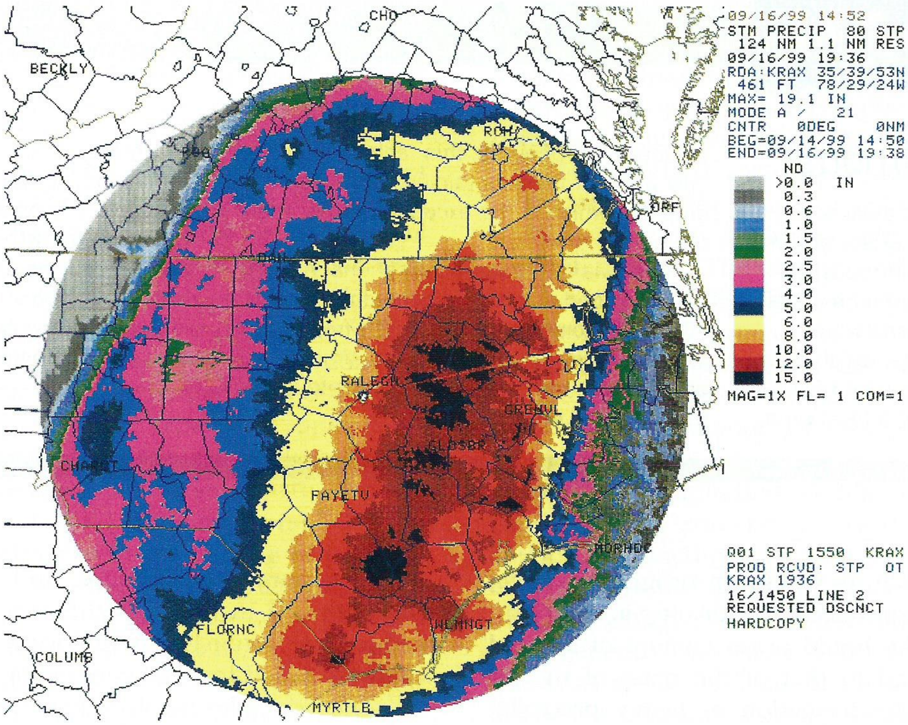
On average, at any moment the atmosphere over the 48 conterminous United States contains the equivalent of 175 trillion liters (40 trillion gallons) of liquid water in the form of water vapor. Slightly more than one-tenth of this moisture condenses and falls to the surface daily as precipitation—enough to yield an average of 76 cm (30 in.) of precipitation annually across the country.

The transmitter/receiver slowly rotates as it emits and receives the radar beams, thus allowing a two-dimensional depiction of the precipitation for several hundred kilometers from the radar. The information can be continually monitored and stored, so that the total amount of precipitation occurring over a fairly extensive region can be estimated (Figure 7–26). Such measurements have proven to be particularly useful in providing short-term forecasting for potential floods.

Snow Measurement

Rain gauges are particularly unreliable when precipitation occurs as snow, because captured snow can obstruct the inlet to the collecting tube. To estimate the precipitation in these environments, we measure the depth of accumulated snow. The *water equivalent* of the snow, which is the depth of water that would result if all the snow melted, can then be roughly estimated using a conversion ratio of 10:1. In reality, the ratio of snow depth to water equivalent can vary greatly—from about 4:1 to 50:1.

In remote mountainous areas, particularly in the western United States and Canada, observations of snow cover have been made for decades at hundreds of *snow courses*. Usually about ten observations are made at each snow course by pushing a collection tube into the snow, extracting the tube and its contents, and weighing them on a spring balance. The weight of the snow-filled tube is directly proportional to the water equivalent of the snow cover, and the average of the ten or so readings is used as the representative value.



▲ **FIGURE 7-26** Precipitation estimates depicted by Doppler radar.

Manual snow course measurements are still obtained but are frequently augmented by automated snow pillows. **Snow pillows** are large air mattresses filled with an antifreeze liquid and connected to pressure recorders. As snow accumulates on a pillow, the increased weight is recorded and converted to a water equivalent. These instruments have radio devices that transmit the data to a centralized receiving station.

Checkpoint

1. Explain how each of the following devices works to measure precipitation: tipping-bucket gauge, weighing-bucket gauge, snow pillow.
2. What are some sources of error in measuring precipitation?

Cloud Seeding

Since the late 1940s, people have tried to induce precipitation from clouds, most often to alleviate droughts. This process, called **cloud seeding**, involves injecting one of two materials into nonprecipitating clouds. The objective is to convert some of the supercooled droplets in a cool cloud to ice and cause precipitation by the Bergeron process.

One of the materials, *dry ice* (frozen carbon dioxide), promotes freezing because of its very low temperature (below −78 °C or −108 °F). Very small shavings of dry ice can be ejected from a plane flying through the cloud. When introduced into a cloud, dry ice lowers the temperature of the droplets so that freezing can occur by homogeneous nucleation (see Chapter 5). (Recall that at temperatures below about −40 °C water droplets require no ice nuclei to freeze.)

The second agent for cloud seeding, *silver iodide*, initiates the Bergeron process by acting as an ice nucleus at temperatures as high as −5 °C (23 °F). Silver iodide owes its effectiveness as an ice nucleus to its six-sided structure. Like dry ice, silver iodide can be introduced directly into a cloud from aircraft. More often, it is mixed with a material that produces smoke when ignited in ground-based burners. Updrafts then carry the smoke and the silver iodide into the cloud. If a portion of the seeded cloud is cold enough, some of the supercooled droplets will freeze and begin the Bergeron process.

The cost-effectiveness of cloud seeding is widely debated. Under ideal circumstances, it can supplement water supplies somewhat. Take, for example, the case of the Sierra Nevada mountain range, which supplies much of the water for California and Nevada. Under the right wind, temperature, and moisture conditions, silver iodide released from the ground can enhance snowfall (which is later released as spring melt) by perhaps 10 percent. The right conditions are not usually met, however. Cloud seeding trials in the mountains of Colorado, Utah, and Montana provided disappointing results.

Strong theoretical reasons provide grounds to doubt the usefulness of cloud seeding, except in regions with a

7–3 FOCUS ON AVIATION

Fog Seeding at Airports

Seeding is sometimes used to clear fogs along airport runways. If a fog exists at temperatures below 0 °C (32 °F), introducing dry ice can instigate the Bergeron process, as it does in clouds. Some of the water droplets freeze into ice crystals, which grow at the expense of water droplets and fall out of the fog as snow, leaving a local area of clear air. Of course, this technique can only work for fogs containing supercooled droplets, which is the exception rather than the norm.

Missoula (Montana) International Airport has had good success seeding fog since 2006. Carbon dioxide is sprayed from a couple of pickup trucks into the air just upwind from the runways to stimulate localized snowfall out of the fog.

Several other methods have been tried for dispersing warm fogs, including flying helicopters over the runways to force down the warm air within a radiation inversion and introducing salt crystals into the fog with the intent of making some droplets larger—thereby accelerating growth by collision and coalescence.

continued uplift of air (such as where an orographic effect exists). Recall that water vapor accounts for only a small portion of the air, and that the liquid water content of a cloud is relatively small compared to that of the mass of the air contained within. Thus, the formation of heavy precipitation requires a constant resupply of moisture into clouds by updrafts. If such updrafts are already occurring, precipitation will probably occur with or without seeding. Consequently, many meteorologists believe that under most circumstances cloud seeding yields little or no additional precipitation.

Another question raises an ethical concern about cloud seeding. Let's assume that seeding a cloud produces rain. Would the cloud have yielded precipitation farther downwind if it had not been seeded? Residents downwind might argue they were deprived of precipitation that would have occurred naturally over their own fields and drainage basins. Such matters have in fact been litigated in civil court. In short, a number of open questions still exist about the value of cloud seeding for enhancing precipitation.

Seeding has also been attempted to reduce hail intensity. It was once believed that seeding hail-producing clouds could increase the number of growing ice pellets. Because clouds contain a limited amount of water that can freeze onto growing hailstones, increasing the number of hailstones would theoretically reduce their average size. The decrease in size would reduce the kinetic energy of the falling hailstones and thereby lower the likelihood of damage near the ground. However, a multiyear experiment in northeastern Colorado in the 1970s failed to support the usefulness of seeding as a hail suppression measure, and such attempts have been discontinued in the United States and Canada. Attempts to reduce the intensity of hurricanes by cloud seeding have likewise failed to yield convincing results.

The principles involved in cloud seeding also apply to the removal of fog. See *Box 7–3, Focus on Aviation: Fog Seeding at Airports,* to learn more about this.

Checkpoint

1. **What are two substances that have been used in cloud seeding?**
2. **Suppose that you are a wheat farmer on the High Plains in Montana. Would cloud seeding intended to save your crop during a drought be cost effective? Explain why or why not.**

Did You Know?

The viability of using dry ice to promote precipitation formation was discovered serendipitously by Vincent Schaefer in 1946. Working with a home freezer in his lab to test whether certain materials could work as ice nuclei, he introduced dry ice into the freezer with moist air to cool it further. He then saw some of the water droplets immediately turn to ice. Further tests by Schaefer and Irving Langmuir proved that introducing dry ice into real clouds could trigger the formation of ice crystals. A few years later, Bernard Vonnegut (brother of novelist Kurt Vonnegut) discovered that silver iodide could also promote ice crystal formation.

Summary

When you first picked up this book, you probably thought you would learn how precipitation occurs. That has been the focus of this chapter.

We can summarize the start of precipitation as the result of cloud droplets growing beyond a size that can remain suspended in the air. The amount of growth necessary for each droplet is tremendous. Some growth occurs through condensation onto existing droplets, but further growth depends on other processes. In the tropics, the primary mechanism for droplet growth is collision and coalescence. Outside the tropics, collision and coalescence are still important, but the Bergeron process dominates, wherein ice crystals grow at the expense of supercooled droplets. Once the Bergeron process has been set in motion, riming and aggregation promote even further growth of ice.

Precipitation occurs in several different forms. Outside the tropics, rain usually results from the melting of ice crystals or snowflakes as they fall to the surface. Ice crystals that do not melt before reaching the surface form snow. Graupel and hail form when supercooled water attaches to ice crystals and freezes. In the case of hail, growth occurs when water repeatedly freezes onto existing ice pellets as they rise above the freezing level on updrafts. Sleet and freezing rain both involve the freezing of raindrops. For sleet, freezing occurs before the drop reaches the surface; for freezing rain, it takes place on contact with the surface.

The standard instrument for measuring precipitation is the simple rain gauge, a collecting device that funnels precipitation into a narrow tube for measurement with a calibrated stick. The timing and intensity of precipitation can be recorded with a modified precipitation gauge called a *tipping-bucket gauge*. As direct and uncomplicated as precipitation measurement may seem, it is subject to a host of potential errors. As a result, our knowledge about the worldwide distribution of precipitation is subject to much uncertainty.

When precipitation is not sufficient to meet human needs, people sometimes resort to cloud seeding, which means introducing materials into clouds to stimulate precipitation by the Bergeron process. This can be done with dry ice, which causes supercooled droplets to freeze by homogeneous nucleation, or with silver iodide, which serves as an ice nucleus. At present, strong reasons exist to doubt the efficacy of cloud seeding in all but very limited circumstances.

Key Terms

drag *page 190*
terminal velocity *page 190*
warm cloud *page 192*
collision–coalescence process *page 192*
collector drop *page 192*
collision *page 192*
coalescence *page 193*
cold cloud *page 193*
cool cloud *page 193*
Bergeron process *page 194*
riming (accretion) *page 195*
aggregation *page 195*
snow *page 196*
lake-effect snow *page 197*
rain *page 199*
shower *page 200*
graupel *page 200*
hail *page 200*
hail cascade *page 201*
sleet *page 202*
freezing rain *page 203*
ice storms *page 203*
rain gauge *page 204*
tipping-bucket gauge *page 204*
weighing-bucket rain gauge *page 204*
snow pillow *page 207*
cloud seeding *page 207*

Review Questions

1. What determines the terminal velocity of falling droplets and raindrops?
2. Describe the characteristics that distinguish warm, cool, and cold clouds.
3. How do the growth processes of droplets in warm and cold clouds differ?
4. Why isn't growth by condensation able to create precipitation-sized droplets on its own?
5. How do collision and coalescence increase the size of cloud droplets?
6. Explain how variations in the saturation vapor pressure for ice crystals and supercooled water droplets affect the development of precipitation.
7. Why can't the Bergeron process take place in warm clouds?
8. What are riming and aggregation?
9. Why is precipitation greater in Mississippi than in Michigan?
10. How do lakes enhance precipitation downwind?
11. Why do rain showers start with only large drops?
12. Explain why the formation of sleet requires an inversion.
13. It is never too cold for snow to occur. Is that also true of sleet?
14. Why does hail consist of multiple layers of ice?
15. What are some inherent sources of error in rain gauges?
16. How do weighing-bucket and tipping-bucket gauges measure rainfall?
17. Explain how snow pillows measure snow accumulation.
18. What materials are used in cloud seeding, and how do they stimulate (or inhibit) precipitation?