anywhere from 500 to 12,000 meters (1650–39,600 ft) above sea level. From this base, they pile up into great rounded structures, often with tops like cauliflowers. The cumulus cloud is the visible evidence of an unstable atmosphere; its base is the point where condensation has begun in a column of air as it moves upward.

Examine Figures 6.10 and 6.11 to familiarize yourself with the basic cloud types and their names. Keep in mind that some cloud shapes exist in all three levels—for example, stratocumulus (strato = low level + cumulus = a rounded shape), altocumulus, and cirrocumulus. These three share the similar rounded or cauliflower appearance of cumulus clouds, which can exist at all three levels. You may notice that altostratus (alto = middle level + stratus = layered shape) and cirrostratus have two-part names, but low-level layered clouds are called stratus only. Lastly, thin, stringy cirrus clouds are found only as high-level clouds, so the term cirro (meaning high-level cloud) is not necessary here.

Other terms used in describing clouds are nimbo or nimbus, meaning precipitation (rain is falling). Thus, the nimbostratus cloud may bring a long-lasting drizzle, and the cumulonimbus is the thunderstorm cloud. This latter cloud has a flat top, called an anvil head, as well as a relatively flat base, and it becomes darker as it grows higher and thicker and thus blocks the incoming sunlight. The cumulonimbus is the source of many atmospheric concerns including high-speed winds, torrential rain, flash flooding, thunder, lightning, hail, and possibly tornadoes. This type of cloud can develop in several different ways as we will soon discuss.

**Adiabatic Heating and Cooling** The cooling process that leads to cloud formation is quite different from that associated with the other condensation forms that we have already examined. The cooling process that produces fog, frost, and dew is either radiation or advection. On the other hand, clouds usually develop from a cooling process that results when a parcel of air on Earth’s surface is lifted into the atmosphere.

The rising parcel of air will expand as it encounters decreasing atmospheric pressure with height. This expansion allows the air molecules to spread out, which causes the parcel’s temperature to decrease. This is known as adiabatic cooling and occurs at the constant lapse rate of approximately 10°C per 1000 meters (5.6°F/1000 ft). By the same token, air descending through the atmosphere is compressed by the increasing pressure and undergoes adiabatic heating of the same magnitude.
However, the rising and cooling parcel of air will eventually reach its dew point—the temperature at which water vapor begins to condense out, forming cloud droplets. From this point on, the adiabatic cooling of the rising parcel will decrease as latent energy released by the condensation process is added to the air. To differentiate between these two adiabatic cooling rates, we refer to the precondensation rate (10°C/1000 m) as the dry adiabatic lapse rate and the lower, postcondensation rate as the wet adiabatic lapse rate. The latter rate averages 5°C per 1000 meters (3.2°F/1000 ft) but varies according to the amount of water vapor that condenses out of the air.

A rising air parcel will cool at one of these two adiabatic rates. Which rate is in operation depends on whether condensation is (wet adiabatic rate) or is not (dry adiabatic rate) occurring.

**FIGURE 6.11**
Types of clouds.
On the other hand, the warming temperatures of descending air allow it to hold greater quantities of water vapor. In other words, as the air temperature rises farther above the dew point, condensation will not occur, so the heat of condensation will not affect the rate of rise in temperature. Thus, the temperature of air that is descending and being compressed always increases at the dry adiabatic rate.

It is important to note that adiabatic temperature changes are the result of changes in volume and do not involve the addition or subtraction of heat from external sources.

It is also extremely important to differentiate between the environmental lapse rate and adiabatic lapse rates. In Chapter 4, we found that in general the temperature of our atmosphere decreases with increasing height above Earth’s surface; this is known as the environmental lapse rate, or the normal lapse rate. Although it averages 6.5°C per 1000 meters (3.6°F/1000 ft), this rate is quite variable and must be measured through the use of meteorological instruments sent aloft. Whereas the environmental lapse rate reflects nothing more than the vertical temperature structure of the atmosphere, the adiabatic lapse rates are concerned with temperature changes as a parcel of air moves through the atmospheric layers (Fig. 6.12).

- FIGURE 6.11 (continued)

- FIGURE 6.12
Comparison of the dry adiabatic lapse rate and the environmental lapse rate. The environmental lapse rate is the average vertical change in temperature. Air displaced upward will cool (at the dry adiabatic rate) because of expansion. In this example, using the environmental lapse rate, what is the temperature of the layer of air at 2000 meters?
Stability and Instability  Although adiabatic cooling results in the development of clouds, the various forms of clouds are related to differing degrees of vertical air movement. Some clouds are associated with rapidly rising, buoyant air, whereas other forms result when air resists vertical movement.

An air parcel will rise of its own accord as long as it is warmer than the surrounding layer of air. When it reaches a layer of the atmosphere that is the same temperature as itself, it will stop rising. Thus, an air parcel warmer than the surrounding atmospheric air will rise and is said to be unstable. On the other hand, an air parcel that is colder than the surrounding atmospheric air will resist any upward movement and will likely sink to lower levels. Then the air is said to be stable.

Determining the stability or instability of an air parcel involves nothing more than asking the question, If an air parcel were lifted to a specific elevation (cooling at an adiabatic lapse rate), would it be warmer, colder, or the same temperature as the atmospheric air (determined by the environmental lapse rate at that time) at that same elevation?

If the air parcel is warmer than the atmospheric air at the selected elevation, then the parcel would be unstable and would continue to rise, because warmer air is less dense and therefore buoyant. Thus, under conditions of instability, the environmental lapse rate must be greater than the adiabatic lapse rate in operation. For example, if the environmental lapse rate is 12°C per 1000 meters and the ground temperature is 30°C, then the atmospheric temperature at 2000 meters would be 6°C. On the other hand, an air parcel (assuming that no condensation occurs) lifted to 2000 meters would have a temperature of 10°C. Because the air parcel is warmer than the atmospheric air around it, it is unstable and will continue to rise (Fig. 6.13).

Now let’s assume that it is another day and all the conditions are the same, except that measurements indicate the environmental lapse rate on this day is 2°C per 1000 meters. Consequently, although our air parcel if lifted to 2000 meters would still have a temperature of 10°C, the temperature of the atmosphere at 2000 meters would now be 26°C. Thus, the air parcel would be colder and would sink back toward Earth as a result of its greater density (see again Fig. 6.13). As you can see, under conditions of stability, the environmental lapse rate is less than the adiabatic lapse rate in operation. If an air parcel, upon being lifted to a specific elevation, has the same temperature as the atmospheric air surrounding it, it is neither stable nor unstable. Instead, it is considered neutral; it will neither rise nor sink but will remain at that elevation.

Whether an air parcel will be stable or unstable is related to the amount of cooling and heating of air at Earth’s surface. With cooling of the air through radiation and conduction on a cool, clear night, air near the surface will be relatively close in temperature to that aloft, and the environmental lapse rate will be low, thus enhancing stability. With the rapid heating of the surface on a hot summer day, there will be a very steep environmental lapse rate because the air near the surface is so much warmer than that above, and instability will be enhanced.

Pressure zones can also be related to atmospheric stability. In areas of high pressure, stability is maintained by the slow subsiding air from aloft. In low pressure regions, on the other hand, instability is promoted by the tendency for air to converge and then rise.

Precipitation Processes

Condensed droplets within cloud formations stay in the air and do not fall to Earth because of their tiny size (0.02 mm, or less than 1000th of an inch), their general buoyancy, and the upward movement of the air within the cloud. These droplets of condensation are so minute that they are kept floating in the cloud formation; their mass and the consequent pull of gravity are insufficient to overcome the buoyant effects of air and the vertical currents, or updrafts, within the clouds. Figure 6.14 shows the relative sizes of a condensation nucleus, a cloud droplet, and a raindrop. It takes about a million cloud droplets to form one raindrop.

Precipitation occurs when the droplets of water, ice, or frozen water vapor grow and develop masses too great to be held aloft. They then fall to Earth as rain, snow, sleet, or hail. The form that precipitation takes depends largely on the method of formation and the temperature during formation. Among the many theories
that try to explain the formation of precipitation, the collision–coalescence process for warm clouds in low latitudes and the Bergeron (or ice crystal) Process for cold clouds at higher latitudes are the most widely accepted.

Precipitation in the lower latitudes of the tropics and in warm clouds is likely to form by the collision–coalescence process. The collision–coalescence process is one in which the name itself describes the process. By nature, water is quite cohesive (able to stick to itself). When water droplets are colliding in the circulation of the cloud, they tend to coalesce (or grow together). This is especially true as the water droplets begin to fall toward the ground. In falling, the larger droplets overtake the smaller, more buoyant droplets and capture them to form even larger raindrops. The mass of these growing raindrops eventually overcomes the updrafts of the cloud and fall to Earth, under the pull of gravity. This process occurs in the warm section of clouds where all the moisture exists as liquid water (Fig. 6.15).

At higher latitudes, storm clouds can possess three distinctive layers. The lowermost is a warm layer of liquid water. Here the temperatures are above the freezing point of 0°C (32°F). Above this is the second layer composed of some ice crystals but mainly supercooled water (liquid water that exists at a temperature below 0°C). In the uppermost layer of these tall clouds, when temperatures are lower than or equal to –40°C (–40°F), ice crystals will dominate (Fig. 6.16). It is in relation to these layered clouds that Scandinavian meteorologist Tor Bergeron presented a more complex explanation.

The Bergeron (or ice crystal) Process begins at great heights in the ice crystal and supercooled water layers of the clouds. Here, the supercooled water has a tendency to freeze on any available surface. (It is for this reason that aircraft flying through middle- to high-latitude thunderstorms run the risk of severe icing and invite disaster.) The ice crystals mixed in with the supercooled water in the highest layers of the clouds can become freezing nuclei and form the centers of growing ice crystals. (Essentially, this is the process that can also create snow.) As the supercooled water continues to freeze onto these frozen nuclei, their masses grow until gravity begins to pull them toward Earth. As this frozen precipitation enters the lower layer of the clouds, the above-freezing temperatures there melt the ice crystals into liquid rain before they hit the ground. Therefore, according to Bergeron, rain in these clouds begins as frozen precipitation and melts into a liquid before reaching Earth.

As the melted precipitation falls through the lower, warmer section of the cloud, the collision–coalescence process may take over and cause the raindrops to grow even larger as they descend toward the surface.

**Major Forms of Precipitation**

Rain, consisting of droplets of liquid water, is by far the most common form of precipitation. Raindrops vary in size but are generally about 2–5 millimeters (approximately 0.1–0.25 in.) in diameter (see again Fig. 6.14). As we all know, rain can come in many ways: as a brief afternoon shower, a steady rainfall, or the deluge of a tropical rainstorm. When the temperature of an air mass is only slightly below the dew point, the raindrops may be very small (about 0.5 mm or less in diameter) and close together. The result is a fine mist called drizzle. Drizzle is so light that it is greatly affected by the direction of air currents and the variability of winds. Consequently, drizzle seldom falls vertically.
Snow is the second most common form of precipitation. When water vapor is frozen directly into a solid without first passing through a stage as liquid water (or sublimation), it forms minute ice crystals around the freezing nuclei (of the Bergeron Process). These crystals characteristically appear as six-sided, symmetric shapes. Combinations of these ice-crystal shapes make up the intricate patterns of snowflakes. Snow will reach the ground if the entire cloud and the air beneath the cloud maintain below-freezing temperatures.

Sleet is frozen rain, formed when rain, in falling to Earth, passes through a relatively thick layer of cold air near the surface and freezes. The result is the creation of small, solid particles of clear or milky ice. In English-speaking countries outside the United States, sleet refers not to this phenomenon of frozen rain but rather to a mixture of rain and snow.

Hail is a less common form of precipitation than the three just described. It occurs most often during the spring and summer months and is the result of thunderstorm activity. Hail appears as rounded lumps of ice, called hailstones, which can vary in size from 5 millimeters (0.2 in.) in diameter and up to sizes larger than a baseball (Fig. 6.17). The world record is a hailstone 30 centimeters (12 in.) in diameter that fell in Australia. Hailstones dropping from the sky can be highly destructive to crops and other vegetation, as well as to cars and buildings. Though primarily a property destroyer, hailstones have been known to kill animals and humans.

Hail forms when ice crystals are lifted by strong updrafts in a cumulonimbus (thunderstorm) cloud. Then, as these ice crystals circulate around the storm cloud, supercooled water droplets attach themselves and are frozen as a layer. Sometimes these pellets are lifted up into the cold layer of air and then dropped again and again. The resulting hailstone, made up of concentric layers of ice, has a frosty, opaque appearance when it finally breaks out of the strong updrafts of the cloud formation and falls to Earth. The larger the hailstone, the more times it is cycled through the freezing process and accumulated additional frozen layers.

On occasion, a raindrop can form and have a temperature below 0°C (32°F). This will occur when there is a shallow layer of below-freezing temperatures all the way to the ground so that the liquid rain can reach a supercooled state. These supercooled droplets will freeze the instant they fall onto a surface that is also at a below-freezing temperature. The resulting icy covering on trees, plants, and telephone and power lines is known as freezing rain (or glaze). People usually call the rain and its blanket of ice an “ice storm” (Fig. 6.18). Because of the weight of ice, glazing can break off large branches of trees, bringing down telephone and power lines. It can also make roads practically impassable. A small counterbalance against the negative effects of glazing is the beauty of the natural landscape after an ice storm. Sunlight catches on the ice, reflecting and making a diamond-like surface covering the most ordinary weeds and tree branches.
The simple explanation

The zones of contact between significant condensation and, ultimately, precipitation. That the heated air rises and thus fulfills the one essential criterion important factor in convection for our discussion of precipitation is denser air around it to complete the convection cycle. The im-

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portant factor in convection for our discussion of precipitation is that the heated air rises and thus fulfills the one essential criterion for significant condensation and, ultimately, precipitation.

To enlarge our understanding of convectional precipitation, let’s apply what we have learned about instability and stability.

Factors Necessary for Precipitation

Three factors are necessary for the formation of any type of precipitation on Earth. The first is the presence of moist air on the surface. This air obviously represents the source of moisture (for the precipitation) and energy (in the form of latent heat of condensation). Second are the condensation nuclei around which the water vapor can condense, discussed earlier in this chapter. Third is a mechanism of uplift. These uplift mechanisms are responsible for forcing the air higher into the atmosphere so that it can cool down (by the dry adiabatic rate) to the dew point. These uplift mechanisms are vital to the process of precipitation.

A parcel of air can be forced to rise in four major ways. All the precipitation that falls anywhere on Earth can be traced back to one of these four uplift mechanisms (Fig. 6.19):

- **Convectional precipitation** results from the displacement of warm air upward in a convectional system.
- **Frontal precipitation** takes place when a warm air mass rises after encountering a colder, denser air mass.
- **Cyclonic (or convergence) precipitation** occurs when air converges upon and is lifted up into a low pressure system.
- **Orographic precipitation** results when a moving air mass encounters a land barrier, usually a mountain, and must rise above it in order to pass.

**Convectional Precipitation** The simple explanation of convection is that when air is heated near the surface it expands, becomes lighter, and rises. It is then displaced by the cooler, denser air around it to complete the convection cycle. The important factor in convection for our discussion of precipitation is that the heated air rises and thus fulfills the one essential criterion for significant condensation and, ultimately, precipitation.

To enlarge our understanding of convectional precipitation, let’s apply what we have learned about instability and stability.

Why are power failures a common occurrence with ice storms?

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Why are power failures a common occurrence with ice storms?

**Frontal Precipitation** The zones of contact between relatively warm and relatively cold bodies of air are known as fronts. When two large bodies of air that differ in density, humidity, and temperature meet, the warmer one is lifted above the colder. When this happens, the major criterion for large-scale condensation and precipitation is once again met. Frontal precipitation thus occurs as the moisture-laden warm air rises above the front caused by contact with the cold air. Continuous frontal precipitation has caused some devastating floods through time.

To fully understand fronts, we must examine what causes unlike bodies of air to come together and what happens when they do. This will be discussed in Chapter 7, where we will take a more detailed look at frontal disturbances and precipitation.
Cyclonic (Convergence) Precipitation

The third mechanism, the cyclonic (also known as convergence), was first introduced in Chapter 5 (see again Fig. 5.4). When air enters a low pressure system, or cyclone, it does so (in a counterclockwise fashion in the Northern Hemisphere) from all directions. When air converges on a low pressure system, it has little option but to rise. Therefore, clouds and possible precipitation are common around the center of a cyclone.

Orographic Precipitation

As was the case with convectional rainfall, orographic rainfall has a simple definition and a somewhat more complex explanation. When land barriers—such as mountain ranges, hilly regions, or even the escarpments (steep edges) of plateaus or tablelands—lie in the path of prevailing winds, large portions of the atmosphere are forced to rise above these barriers. This fills the one main criterion for significant precipitation—that large masses of air are cooled by ascent and expansion until large-scale condensation takes place. The resultant precipitation is termed orographic (from Greek: oros, mountains). As long as the air parcel rising up the mountainside remains stable (cooling at a greater rate than the environmental lapse rate), any resulting cloud cover will be a type of stratus cloud. However, the situation can be complicated by the same circumstances illustrated in Figure 6.20b. A potentially unstable air parcel may need only the initial lift provided by the orographic barrier to set it in motion. In this case, it will continue to rise of its own accord (no longer forced) as it seeks air of its own temperature and density. Once the land barrier provides the initial thrust, it has performed its function as a lifting mechanism.

Because the air deposits most of its moisture on the windward side of a mountain, there will normally be a great deal less precipitation on the leeward side; on this side, the air will be much drier and the dew point consequently much lower. Also, as air descends the leeward slope, its temperature warms (at the dry adiabatic rate), and condensation ceases. The leeward side of the mountain is said to be in the rain shadow (Fig. 6.21a). Just as being in the shade, or in shadow, means that you are not receiving any direct sun, so being in the rain shadow means that you do not receive much rain. If you live near a mountain range, you can see the effects of orographic precipitation and the rain shadow in the pattern of vegetation (Figs. 6.21b and c). The windward side of the mountains (say, the Sierra Nevada in California) will be heavily forested and thick with vegetation. The opposite slopes in the rain shadow will usually be drier and the cover of vegetation sparser.

Distribution of Precipitation

The precipitation a region receives can be described in different ways. We can look at average annual precipitation to get an overall picture of the amount of moisture that a region gets during a year. We can also look at its number of raindays—days on which 1.0 millimeter (0.01 in.) or more of rain is received during a 24-hour period. Less than this amount is known as a trace of rain. If we divide the number of raindays in a month or year by the total number of days in that period, the resulting figure represents the probability of rain. Such a measure is important to farmers and to ski or summer resort owners whose incomes may depend on precipitation or the lack of it.

We can also look at the average monthly precipitation. This provides a picture of the seasonal variations in precipitation (Fig. 6.22). For instance, in describing the climate of the west...
coast of California, average annual precipitation would not give the full story because this figure would not show the distinct wet and dry seasons that characterize this region.

**Horizontal Distribution of Precipitation**  
Figure 6.23 shows average annual precipitation for the world’s continents. We can see that there is great variability in the distribution of precipitation over Earth’s surface. Although there is a zonal distribution of precipitation related to latitude, this distribution is obviously not the only factor involved in the amount of precipitation an area receives.

The likelihood and amount of precipitation are based on two factors. First, precipitation depends on the degree of lifting that occurs in air of a particular region. This lifting, as we have already seen, may be due to the collision of different air masses (frontal), to the convergence of air into a low pressure system (cycloonic or convergence), to differential heating of Earth’s surface (convection), to the lifting that results when an air mass encounters a rise in Earth’s surface (orographic), or to a combination of these processes. The second factor affecting the likelihood of precipitation depends on the internal characteristics of the air itself, including its degree of instability, its temperature, and its humidity.

Because higher temperatures, as we have seen, allow air masses to hold greater amounts of water vapor and because, conversely, cold air masses can hold less water vapor, we can expect a general decrease of precipitation from the equator to the poles that is related to the unequal zonal distribution of incoming solar energy discussed in Chapter 3.

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**FIGURE 6.20**  
Effect of humidity on air mass stability. (a) Warm, dry air rises and cools at the dry adiabatic rate, soon becoming the same temperature as the surrounding air, at which point convectional uplift terminates. Because the rising dry air did not cool to its dew point temperature by the time that convectional lifting ended, no cloud formed. (b) Rising warm, moist air soon cools to its dew point temperature. The upward-moving air subsequently cools at the wet adiabatic rate, which keeps the air warmer than the surrounding atmosphere so that the uplift continues. Only when all moisture is removed by condensation will the air cool rapidly enough at the dry adiabatic rate to become stable.

What would be necessary for the cloud in (b) to stop its upward growth at 4500 meters?
However, if we look again at Figure 6.23, we see a great deal of variability in average annual precipitation beyond the general pattern of a decrease with increased latitude. In the following discussion, we examine some of these variations and give the reasons for them. We also apply what we have already learned about temperature, pressure systems, wind belts, and precipitation.

Distribution within Latitudinal Zones The equatorial zone is generally an area of high precipitation—more than 200 centimeters (79 in.) annually—largely due to the zone’s high temperatures, high humidity, and the instability of its air. High temperatures and instability lead to a general pattern of rising air, which in turn allows for precipitation. This tendency is strongly reinforced by the convergence of the trades as they move toward the equator from opposite hemispheres. In fact, the intertropical convergence zone is one of the two great zones where air masses converge. (The other is along the polar front within the westerlies.)

In general, the air of the trade wind zones is stable compared with the instability of the equatorial zone. Under the control of these steady winds, there is little in the way of atmospheric disturbances to lead to convergent or convectional lifting. However, because the trade winds are basically easterly, when they

![Diagram](image-url)

**FIGURE 6.21**
Orographic precipitation and the rain-shadow effect. (a) Orographic uplift over the windward (western) slope of the Sierras produces condensation, cloud formation, and precipitation, resulting in (b) dense stands of forest. (c) Semiarid or rain-shadow conditions occur on the leeward (eastern) slope of the Sierras.

Can you identify a mountain range in Eurasia in which the leeward side of that range is in the rain shadow?

![Table](image-url)

**FIGURE 6.22**
Average monthly precipitation in San Francisco, California, is represented by colored bars along the bottom of the graph. A graph of monthly precipitation figures like this one gives a much more accurate picture than the annual precipitation total, which does not tell us that nearly all the precipitation occurs in only half of the year.

How would this rainfall pattern affect agriculture?
When you look at clouds in our atmosphere, it is often quite easy to see their relatively flat bases. Cloud tops may appear quite irregular, but cloud bases are often flat. Even if the cloud bases do not seem flat, it will be obvious that the clouds you see all seem to be formed at the same level above the surface. This level represents the altitude to which the air must be lifted (and cooled at the dry adiabatic rate) before saturation is reached. Any additional lifting and clouds will form and build upward. Therefore, the height at which clouds form from lifting is called the lifting condensation level (LCL) and can be estimated by the equation:

\[ \text{LCL (in meters)} = 125 \times (\text{Celsius temperature} - \text{Celsius dew point}) \]

For example, if the surface temperature is 7.2°C (45°F) and the dew point temperature is 4.4°C (40°F), then the LCL is estimated at 350 meters (1148 ft) above the surface.

Caution: Keep in mind that different layers of clouds may exist at the same time. Low, middle, and high clouds as defined in this chapter may all appear on the same afternoon. These clouds may have formed in other regions and be only passing overhead. The formula presented here is best used with the lowest level of cloud cover that appears overhead.

The stratocumulus clouds (bottom layer) show the lifting condensation level (LCL).