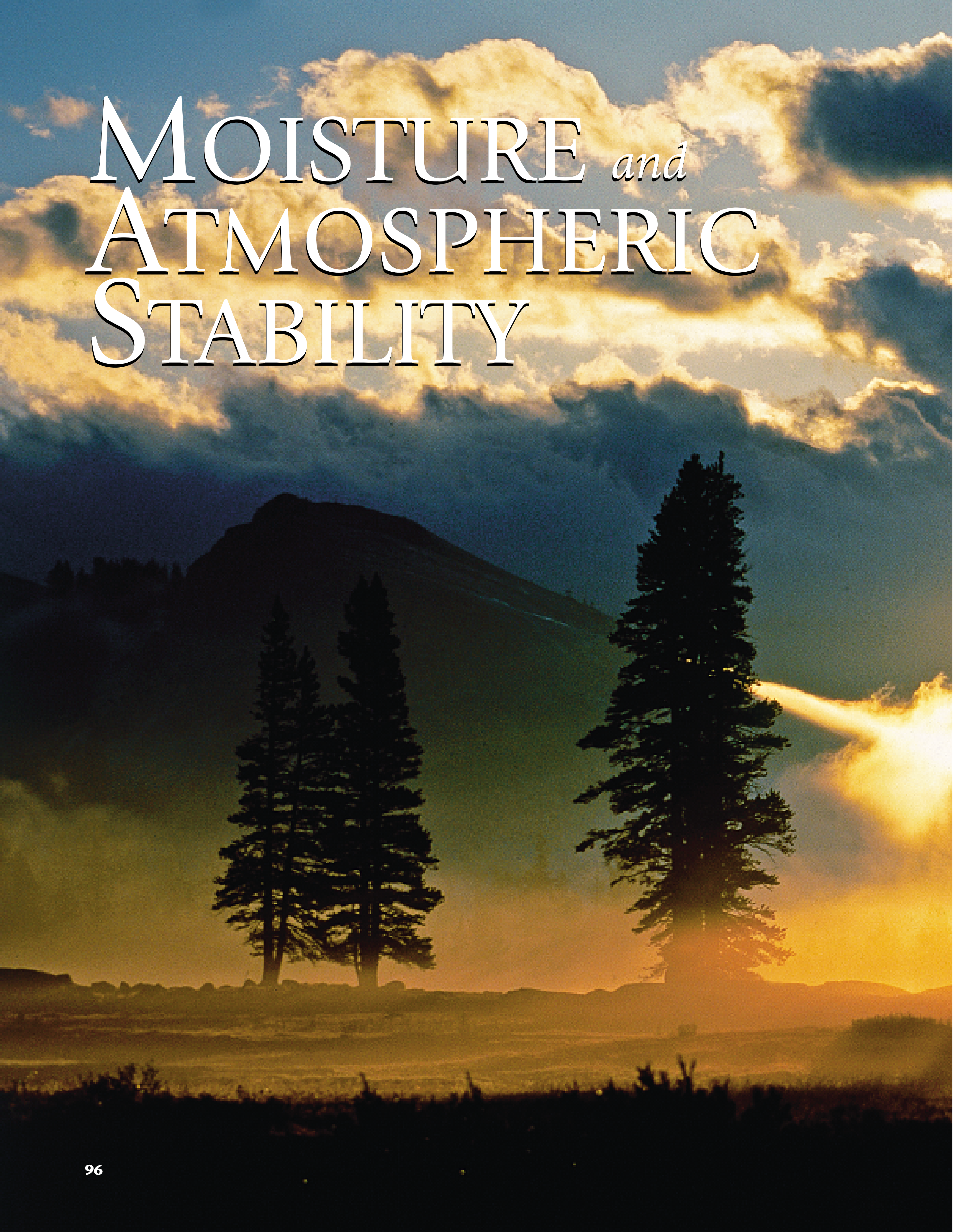
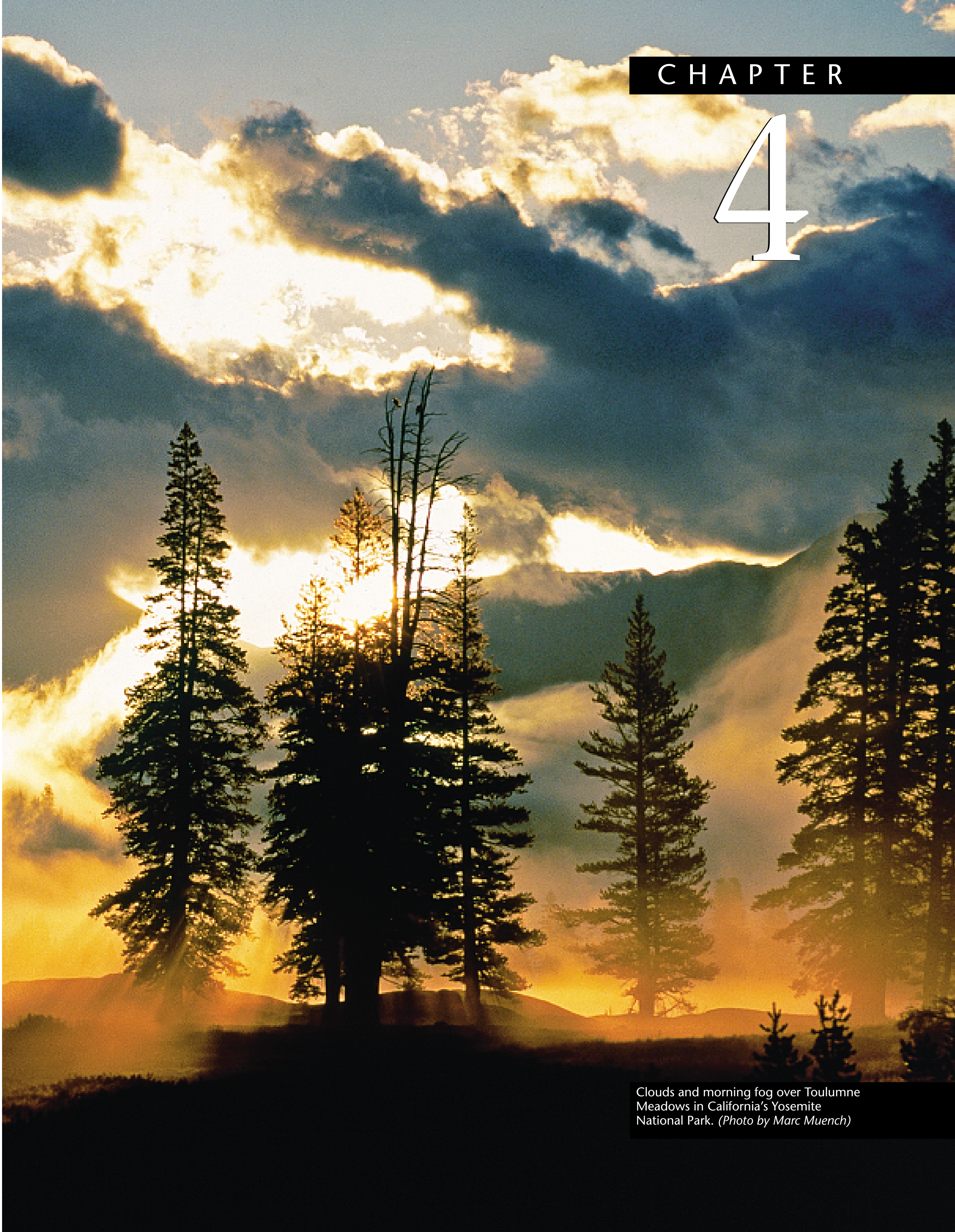
MOISTURE *and* ATMOSPHERIC STABILITY



CHAPTER 4



Clouds and morning fog over Toulumne Meadows in California's Yosemite National Park. *(Photo by Marc Muench)*

Water vapor is an odorless, colorless gas that mixes freely with the other gases of the atmosphere. Unlike oxygen and nitrogen—the two most abundant components of the atmosphere—water can change from one state of matter to another (solid, liquid, or gas) at the temperatures and pressures experienced on Earth. (By contrast, nitrogen will not condense to a liquid unless its temperature is lowered to −196°C [−371°F]). Because of this unique property, water freely leaves the oceans as a gas and returns again as a liquid.

As you observe day-to-day weather changes, you might ask: Why is it generally more humid in the summer than in the winter? Why do clouds form on some occasions but not on others? Why do some clouds look thin and harmless, whereas others form gray and ominous towers? Answers to these questions involve the role of water vapor in the atmosphere, the central theme of this chapter.

Movement of Water Through the Atmosphere

GEODe ATMOSPHERE

Moisture and Cloud Formation

▸ Movement of Water Through the Atmosphere

Water is everywhere on Earth—in the oceans, glaciers, rivers, lakes, the air, soil, and in living tissue (Figure 4–1). All of these "reservoirs" constitute Earth's hydrosphere. In all, the water content of Earth's hydrosphere is about 1.36 billion cubic kilometers (326 million cubic miles).

The increasing demands on this finite resource have led scientists to focus on the continuous exchange of water among the oceans, the atmosphere, and the continents (Figure 4–2). This unending circulation of Earth's water supply has come to be called the **hydrologic cycle** (or water cycle).

The hydrologic cycle is a gigantic system powered by energy from the Sun in which the atmosphere provides the vital link between the oceans and continents. Water from the oceans and, to a much lesser extent, from the continents, evaporates into the atmosphere. Winds transport this moisture-laden air, often over great distances.

Complex processes of cloud formation eventually result in precipitation. The precipitation that falls into the ocean has ended its cycle and is ready to begin another by evaporating again. The water that falls on the continents, however, must still flow back to the oceans.

Once precipitation has fallen on land, a portion of the water soaks into the ground, some of it moving downward, then laterally, and finally seeping into lakes and streams or directly into the ocean. Much of the water that soaks in or runs off eventually finds its way back to the atmosphere. In addition to evaporation from the soil, lakes, and streams, some water that infiltrates the ground is absorbed by plants through their roots. They then release it into the atmosphere, a process called **transpiration.**

FIGURE 4-1 Steam fog over a lake in Maine. *(Photo by Sara Gray/Getty Image Inc.-Stone Allstock)*

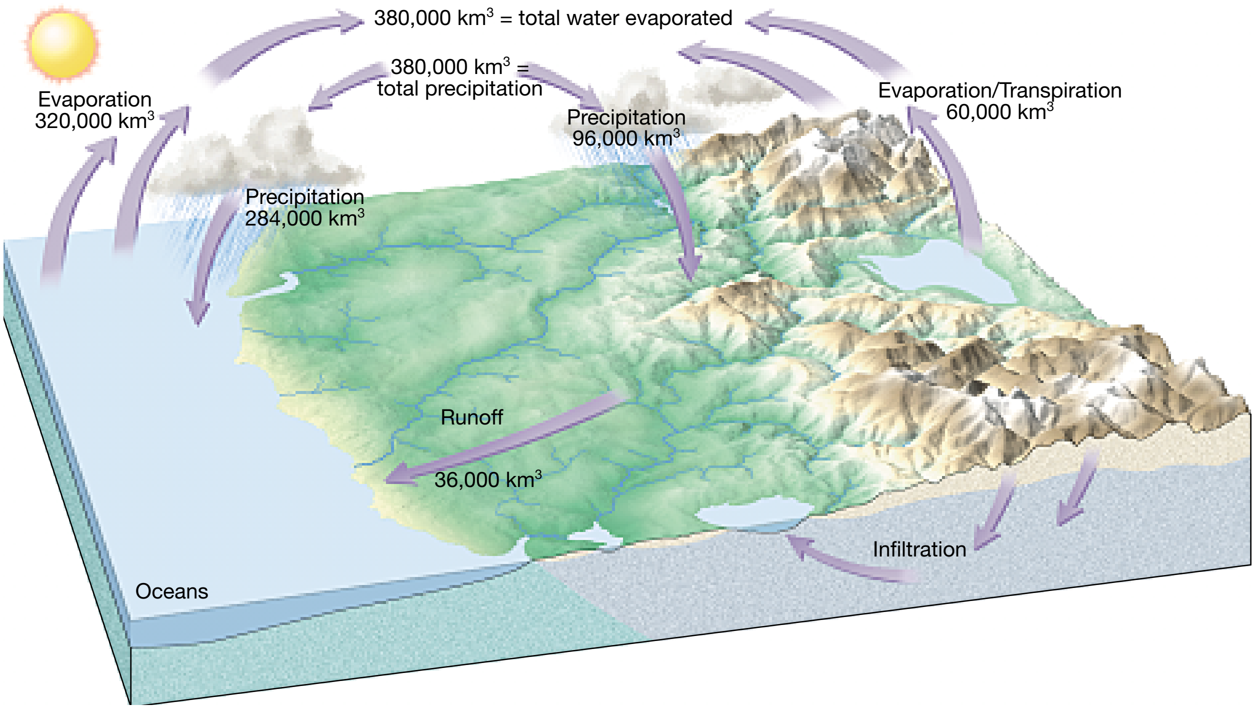


FIGURE 4-2 Earth's water balance. About 320,000 cubic kilometers of water are evaporated annually from the oceans, whereas evaporation from the land (including lakes and streams) contributes 60,000 cubic kilometers of water. Of this total of 380,000 cubic kilometers of water, about 284,000 cubic kilometers fall back to the ocean, and the remaining 96,000 cubic kilometers fall on Earth's land surface. Because 60,000 cubic kilometers of water leave the land through evaporation and transpiration, 36,000 cubic kilometers of water remain to erode the land during the journey back to the oceans.

Figure 4–2 not only shows Earth's hydrologic cycle but also its *water balance*. The water balance is a quantitative view of the hydrologic cycle. Although the amount of water vapor in the air is just a tiny fraction of Earth's total water supply, the absolute quantities that are cycled through the atmosphere in a year are immense, some 380,000 cubic kilometers (91,000 cubic miles). This is enough to cover Earth's surface uniformly to a depth of about 1 meter (3.3 feet). Estimates show that over North America almost six times more water is carried within the moving currents of air than is transported by all the continent's rivers.

Because the total amount of water vapor in the entire global atmosphere remains about the same, the average annual precipitation over Earth must be equal to the quantity of water evaporated. However, for the continents, precipitation exceeds evaporation. Conversely, over the oceans, evaporation exceeds precipitation. Because the level of the world ocean is consistent, runoff from land areas must balance the deficit of precipitation over the oceans.

In summary, the hydrologic cycle depicts the continuous movement of water from the oceans to the atmosphere, from the atmosphere to the land, and from the land back to the sea. The movement of water through the cycle holds the key to the distribution of moisture over the surface of our planet and is intricately related to all atmospheric phenomena.

Water's Changes of State

GEODe ATMOSPHERE **Moisture and Cloud Formation**
▶ Water's Changes of State

Water is the only substance that exists in the atmosphere as a solid (ice), liquid, and gas (water vapor). It is made of hydrogen and oxygen atoms that are bonded together to form water molecules (H_2O). In all three states of matter (even ice) these molecules are in constant motion—the higher the temperature, the more vigorous the movement. The chief difference among liquid water, ice, and water vapor is the arrangement of the water molecules.

Ice, Liquid Water, and Water Vapor

Ice is composed of water molecules that have low kinetic energies (motion) and are held together by mutual molecular attractions (see Box 4–1). Here the molecules form a tight, orderly network as shown in Figure 4–3. As a consequence, the water molecules in ice are not free to move relative to each other but rather vibrate about fixed sites. When ice is heated, the molecules oscillate more rapidly. When the rate of molecular movement increases sufficiently, the bonds between some of the water molecules are broken, resulting in melting.

BOX 4-1 Water: A Unique Substance

Water is the only *liquid* found at Earth's surface in large quantities. Water has several unusual properties that set it apart from other substances, for instance: 1) water is readily converted from one state of matter to another (solid, liquid, gas); 2) water's solid phase, ice, is less dense than liquid water; and 3) water has an unusually high heat capacity. All of these properties influence Earth's weather and climate and are favorable to life as we know it.

These unusual properties are largely a consequence of water's ability to form hydrogen bonds. To better grasp the nature of hydrogen bonds, let's examine a water molecule. A water molecule (H_2O) consists of two hydrogen atoms that are strongly bonded to an oxygen atom. Because oxygen atoms have a greater affinity for the bonding electrons (negatively charged subatomic particles) than do hydrogen atoms, the oxygen end of a water molecule acquires a partial negative charge. For the same reason, both hydrogen atoms on a water molecule acquire a partial positive charge. Because oppositely charged particles attract, a hydrogen atom on one water molecule is attracted to an oxygen atom on another water molecule. Thus, *hydrogen bonding* in water is an attractive force that exists between a hydrogen atom in one water molecule and an oxygen atom of any other water molecule.

Hydrogen bonds are what hold water molecules together to form the solid we call *ice*. In ice, hydrogen atoms form hydrogen bonds with oxygen atoms in other water molecules, producing the hexagonal network shown in Figure 4–A. The resulting molecular configuration is very open (lots of empty space), which accounts for ice being less dense than liquid water. It is important to note that the attractive forces between hydrogen and oxygen that form water molecules (H_2O) are much stronger than the hydrogen bonds that form ice. Thus, when ice is heated, it is only the hydrogen bonds that are broken.

When ice melts, some of the hydrogen bonds (but not all) are broken. This allows for the more compact arrangement shown in Figure 4–B. Consequently, the liquid phase of water is denser than the solid phase. Should water's temperature

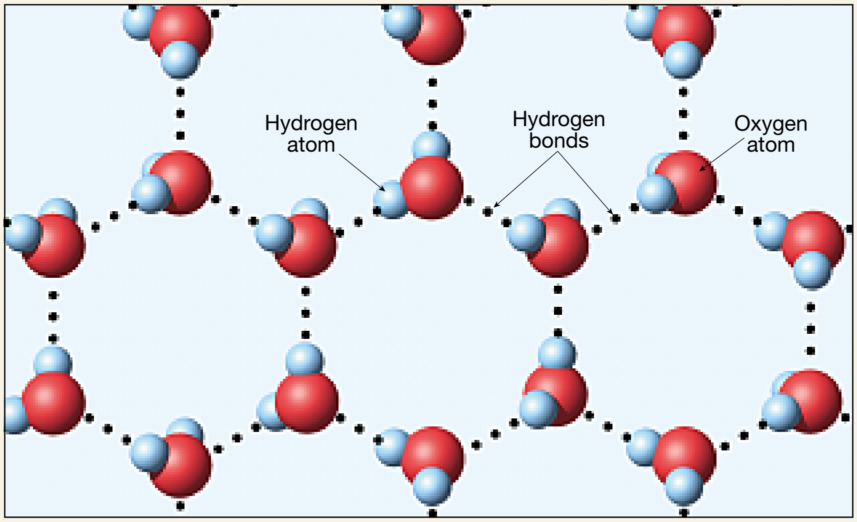


FIGURE 4-A Illustration of the crystalline structure of ice. Water molecules are joined together by hydrogen bonds (shown with dotted lines) that connect a hydrogen atom on one water molecule with an oxygen atom on another. Oxygen atoms are represented by red spheres and hydrogen atoms by small blue spheres. For simplicity the three-dimensional nature of ice has not been illustrated.

In the liquid state, water molecules are still tightly packed but are moving fast enough that they are able to easily slide past one another. As a result, liquid water is fluid and will take the shape of its container.

As liquid water gains heat from its environment, some of the molecules will acquire enough energy to break the remaining molecular attractions and escape from the surface, becoming water vapor. Water-vapor molecules are widely spaced compared to liquid water and exhibit very energetic random motion. What distinguishes a gas from a liquid is its compressibility (and expandability). For example, you can easily put more and more air into a tire and increase its volume only slightly. However, don't try to put 10 gallons of gasoline into a five-gallon can.

To summarize, when water changes state, it does not turn into a different substance; only the distances and interactions among the water molecules changes.

Latent Heat

Whenever water changes state, heat is exchanged between water and its surroundings. When water evaporates, heat is absorbed (Figure 4–3). Meteorologists often measure heat

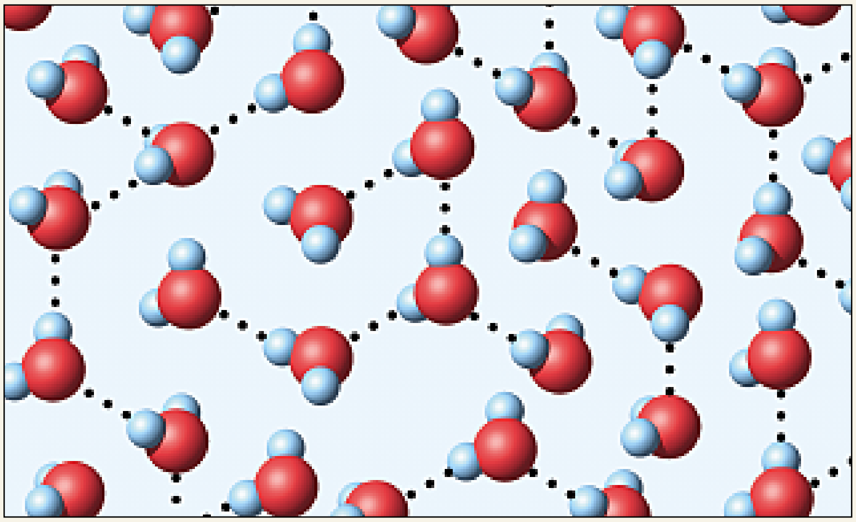


FIGURE 4-B Water in the liquid state consists of clusters of water molecules joined together by hydrogen bonds. As these clusters move about, they continually break up and are replaced by new ones.

rise, more hydrogen bonds will break, and the liquid will become even denser, until it reaches 4°C (39°F) when maximum density is achieved. At higher temperatures water gradually becomes less dense because of increased molecular motion—just like most other substances. However, even at higher temperatures, liquid water has significant hydrogen bonding that tends to form clusters of water molecules that are free to move relative to each other. These clusters continually break up and are replaced by new ones.

As a body of water freezes, ice forms on top because ice is less dense than the water beneath it. This has far-reaching effects, both for our daily weather and for aquatic life. When ice forms on a water body, it insulates the underlying liquid and limits further freezing. If a water body froze from the bottom, imagine the consequences. Many lakes would freeze solid during the winter, killing the aquatic life. In addition, deep bodies of water, such as the Arctic Ocean, would never be ice-covered. This would alter Earth's heat budget, which in turn would modify global atmospheric and oceanic circulations.

Water's heat capacity is also related to hydrogen bonding. When water is heated, some of the energy is used to break hydrogen bonds rather than to increase molecular motion. (Recall that an increase in average molecular motion corresponds to an increase in temperature.) Thus, under similar conditions, water heats up and cools down more slowly than most common substances. This contributes to the differential heating of land and water discussed in Chapter 3. Large water bodies tend to moderate temperatures by remaining warmer than adjacent landmasses in winter and remaining cooler in summer.

Hydrogen bonds are also broken as water evaporates. Thus, water molecules need "extra energy" to give them the motion needed to break the molecular attractions and escape the surface of the liquid to become a gas. (When liquids evaporate, the energy absorbed by the gas molecules is called *latent heat of vaporization.*) Water vapor molecules contain an unusually large amount of "stored energy," which is an important factor influencing weather and climate. When water vapor condenses in the atmosphere, this energy is released, warming the surrounding air and giving it buoyancy. If the moisture content of air is high, this process can spur the development of thunderstorms and contribute to hurricane formation in the tropics.

energy in calories. One **calorie** is the amount of heat required to raise the temperature of 1 gram of water 1°C (1.8°F). Thus, when 10 calories of heat are absorbed by 1 gram of water, the molecules vibrate faster and a 10°C (18°F) temperature rise occurs.

Under certain conditions, heat may be added to a substance without an accompanying rise in temperature. For example, when a glass of ice water is warmed, the temperature of the ice-water mixture remains a constant 0°C (32°F) until all the ice has melted. If adding heat does not raise the temperature, where does this energy go? In this case, the added energy went to break the molecular attractions that bind water molecules into a crystalline structure.

Because the heat used to melt ice does not produce a temperature change, it is referred to as **latent heat.** (Latent means *hidden,* like the latent fingerprints hidden at a crime scene.) This energy can be thought of as being stored in liquid water, and it is not released to its surroundings as heat until the liquid returns to the solid state.

It requires 80 calories to melt one gram of ice, an amount referred to as *latent heat of melting. Freezing,* the reverse process, releases these 80 calories per gram to the

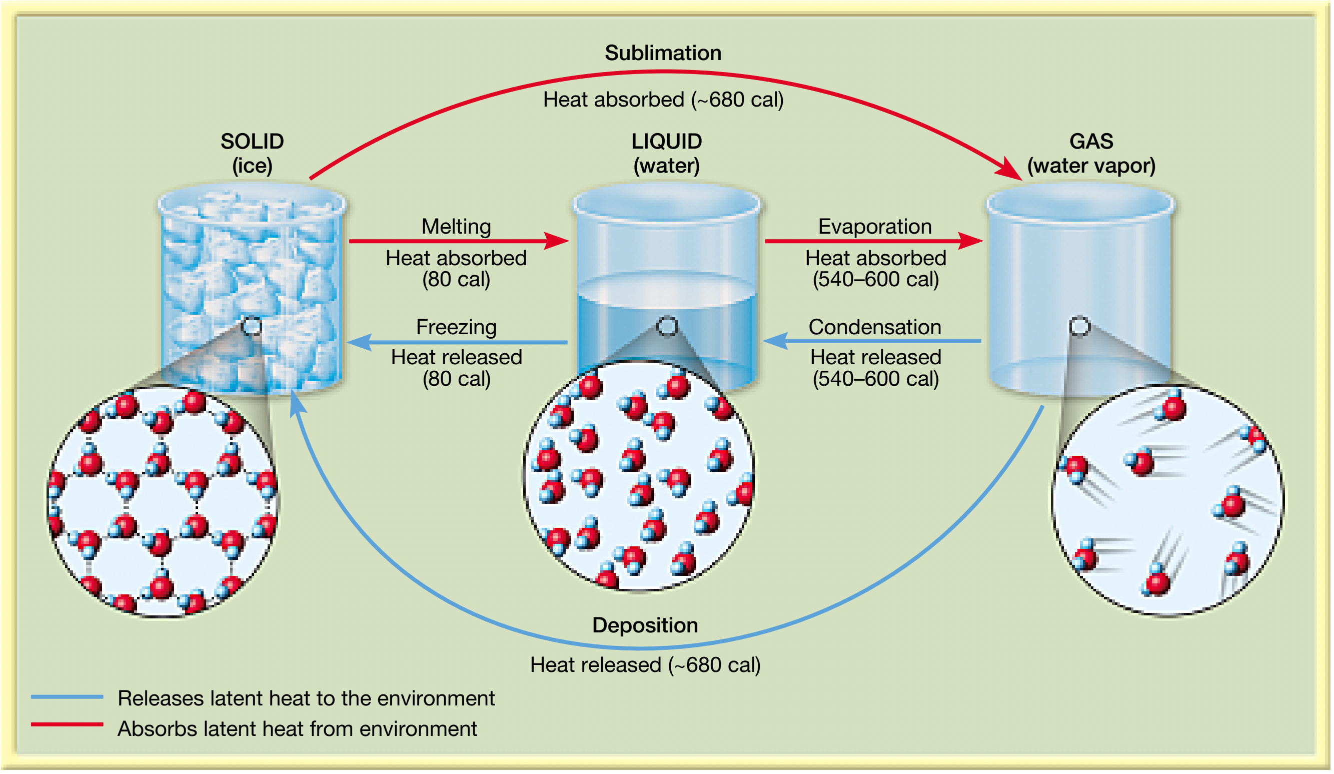


FIGURE 4-3 Change of state always involves an exchange of heat. The amounts of heat are expressed in calories and are shown here for the change of 1 gram of water from one state of matter to another.

environment as *latent heat of fusion*. We will consider the importance of latent heat of fusion in Chapter 5 in the section on frost prevention.

Evaporation and Condensation. We saw that heat is absorbed when ice is converted to liquid water. Heat is also absorbed during **evaporation,** the process of converting a liquid to a gas (vapor). The energy absorbed by water molecules during evaporation is used to give them the motion needed to escape the surface of the liquid and become a gas. This energy is referred to as the **latent heat of vaporization** and varies from about 600 calories per gram for water at 0°C to 540 calories per gram at 100°C. (Notice from Figure 4–3 that it takes much more energy to evaporate 1 gram of water than it does to melt the same amount of ice.) During the process of evaporation, it is the higher-temperature (faster-moving) molecules that escape the surface. As a result, the average molecular motion (temperature) of the remaining water is reduced—hence, the common expression "Evaporation is a cooling process." You have undoubtedly experienced this cooling effect on stepping dripping wet from a swimming pool or bathtub. In this situation the energy used to evaporate water comes from your skin—hence, you feel cool.

Condensation, the reverse process, occurs when water vapor changes to the liquid state. During condensation, water-vapor molecules release energy (**latent heat of condensation**) in an amount equivalent to what was absorbed during evaporation. When condensation occurs in the atmosphere, it results in the formation of such phenomena as fog and clouds (Figure 4–4).

As you will see, latent heat plays an important role in many atmospheric processes. In particular, when water vapor condenses to form cloud droplets, latent heat of condensation is released, warming the surrounding air and giving it buoyancy. When the moisture content of air is high, this process can spur the growth of towering storm clouds. Furthermore, the evaporation of water over the tropical oceans and the subsequent condensation at higher latitudes results in significant energy transfer from equatorial to more poleward locations.

Sublimation and Deposition. You are probably least familiar with the last two processes illustrated in Figure 4–3—sublimation and deposition. **Sublimation** is the conversion of a solid directly to a gas without passing through the liquid state. Examples you may have observed include the gradual shrinking of unused ice cubes in the freezer and the rapid conversion of dry ice (frozen carbon dioxide) to wispy clouds that quickly disappear.

Deposition refers to the reverse process, the conversion of a vapor directly to a solid. This change occurs, for example, when water vapor is deposited as ice on solid objects such as grass or windows (Figure 4–5). These deposits are called *white frost* or *hoar frost* and are frequently referred to simply as *frost*. A household example of the process of deposition is the "frost" that accumulates in a freezer. As shown in Figure 4–3, deposition releases an amount of energy equal to the total amount released by condensation and freezing.

FIGURE 4-4 Condensation of water vapor generates phenomena such as clouds and fog. *(Photo by Jeremy Walker/Getty Image Inc.-Stone Allstock)*

Students Sometimes Ask

What is "freezer burn"?

Freezer burn is what happens to poorly wrapped food stored in a frost-free refrigerator for a few months. Because frost-free refrigerators are designed to remove moisture from the freezer compartment, the air within them is relatively dry. As a result, the moisture in food sublimates—turns from ice to water vapor—and escapes. Thus, the food is not actually burned; it is simply dried out.

FIGURE 4-5 White frost on a window pane. *(Photo by Craig F. Bohren)*

Humidity: Water Vapor in the Air

GEODe ATMOSPHERE

Moisture and Cloud Formation

▶ Humidity: Water Vapor in the Air

Water vapor constitutes only a small fraction of the atmosphere, varying from as little as one-tenth of 1 percent up to about 4 percent by volume. But the importance of water in the air is far greater than these small percentages would indicate. Indeed, scientists agree that *water vapor* is the most important gas in the atmosphere when it comes to understanding atmospheric processes.

Humidity is the general term used to describe the amount of water vapor in the air Figure 4–6. Meteorologists employ several methods to express the water-vapor content of the air, including (1) absolute humidity, (2) mixing ratio, (3) vapor pressure, (4) relative humidity, and (5) dew point. Two of these methods, *absolute humidity* and *mixing ratio*, are similar in that both are expressed as the quantity of water vapor contained in a specific amount of air.

Absolute humidity is *the mass of water vapor in a given volume of air* (usually as grams per cubic meter).

$$\text{Absolute humidity} = \frac{\text{mass of water vapor (grams)}}{\text{volume of air (cubic meters)}}$$

As air moves from one place to another, changes in pressure and temperature cause changes in its volume. When such volume changes occur, the absolute humidity also changes, even if no water vapor is added or removed. Consequently, it is difficult to monitor the water-vapor content of a moving mass of air if absolute humidity is the index

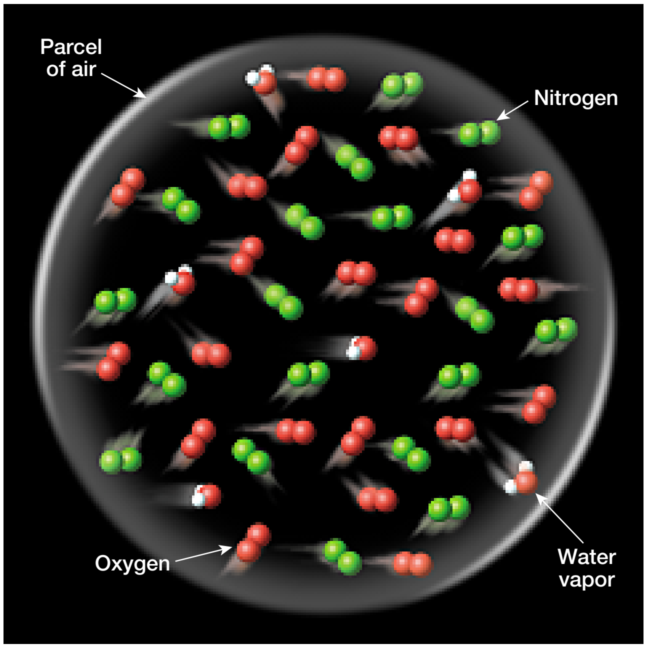


FIGURE 4-6 Meteorologists use several methods to express the water-vapor content of air.

being used. Therefore, meteorologists generally prefer to employ mixing ratio to express the water-vapor content of air.

The **mixing ratio** is *the mass of water vapor in a unit of air compared to the remaining mass of dry air.*

$$\text{mixing ratio} = \frac{\text{mass of water vapor (grams)}}{\text{mass of dry air (kilograms)}}$$

Because it is measured in units of mass (usually grams per kilogram), the mixing ratio is not affected by changes in pressure or temperature. (Figure 4–7).*

Neither the absolute humidity nor the mixing ratio, however, can be easily determined by direct sampling. Therefore, other methods are also used to express the moisture content of the air. These include vapor pressure, relative humidity, and dew point.

Vapor Pressure and Saturation

Another measure of the moisture content of the air is obtained from the pressure exerted by water vapor. To understand how water vapor exerts pressure imagine a closed container half full of pure water and overlain by dry air, as shown in Figure 4–8a. Almost immediately some of the water molecules begin to leave the water surface and evaporate into the dry air above. The addition of water vapor into the air can be detected by a small increase in pressure

*Another commonly used expression is "specific humidity," which is the mass of water vapor in a unit mass of air, including the water vapor. Because the amount of water vapor in the air rarely exceeds a few percent of the total mass of the air, the specific humidity of air is equivalent to its mixing ratio for all practical purposes.

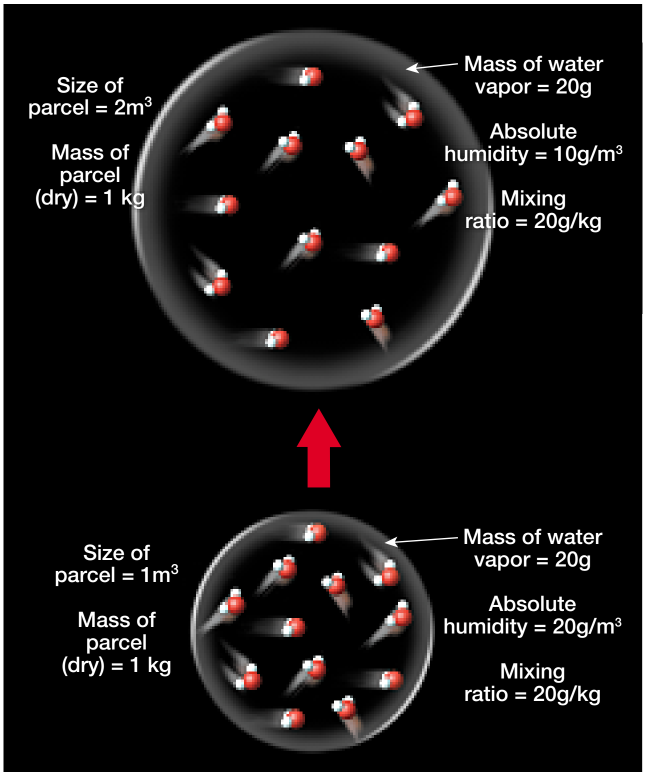


FIGURE 4-7 Comparison of absolute humidity and mixing ratio for a rising parcel of air. Note that the mixing ratio is not affected by changes in pressure as the parcel of air rises and expands.

(Figure 4–8b). This increase in pressure is the result of the motion of the water-vapor molecules that were added to the air through evaporation. In the atmosphere, this pressure is called **vapor pressure** and is defined as *that part of the total atmospheric pressure attributable to its water-vapor content.*

Initially, many more molecules will leave the water surface (evaporate) than will return (condense). However, as more and more molecules evaporate from the water surface, the steadily increasing vapor pressure in the air above forces more and more water molecules to return to the liquid. Eventually a balance is reached in which the number of water molecules returning to the surface balances the number leaving. At that point the air is said to have reached an equilibrium called **saturation** (Figure 4–8c). When air is saturated, the pressure exerted by the motion of the water-vapor molecules is called the **saturation vapor pressure.**

Now suppose we were to disrupt the equilibrium by heating the water in our closed container, as illustrated in Figure 4–8d. The added energy would increase the rate at which the water molecules would evaporate from the surface. This in turn would cause the vapor pressure in the air above to increase until a new equilibrium was reached between evaporation and condensation. Thus, we conclude that the saturation vapor pressure is temperature-dependent, such that at higher temperatures it takes more water vapor to saturate air (Figure 4–9). The amount of

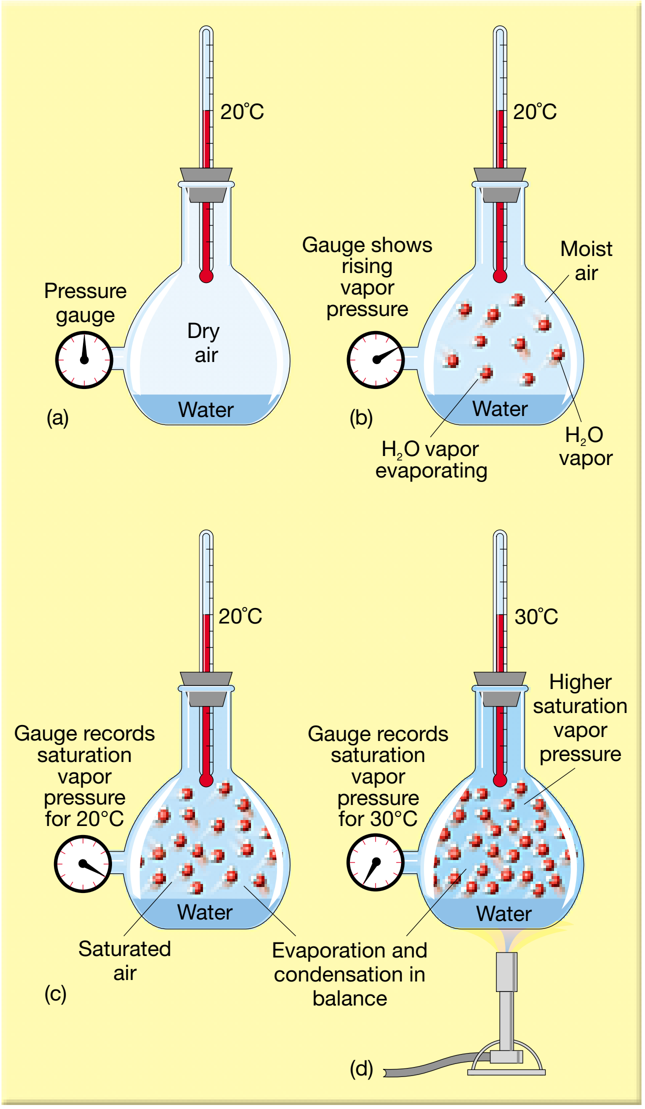


FIGURE 4-8 The relationship between vapor pressure and saturation. (a) Initial conditions—dry air at 20°C with no observable vapor pressure. (b) Evaporation generates measurable vapor pressure. (c) As more and more molecules escape from the water surface, the steadily increasing vapor pressure forces more and more of these molecules to return to the liquid. Eventually, the number of water-vapor molecules returning to the surface will balance the number leaving. At that point the air is said to be saturated. (d) When the container is heated from 20° to 30°C, the rate of evaporation increases, causing the vapor pressure to increase until a new balance is reached.

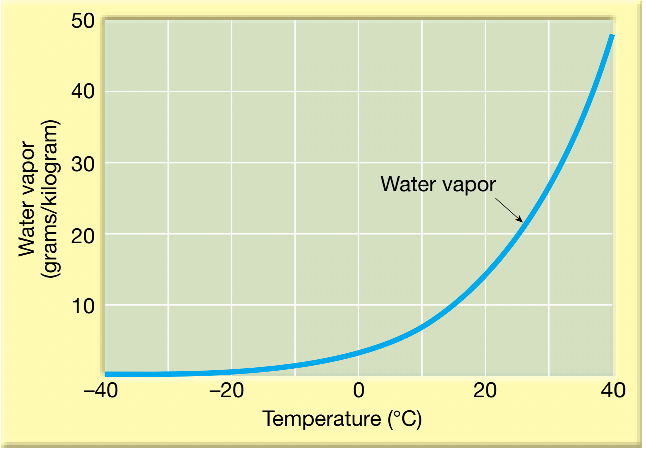


FIGURE 4-9 The amount of water vapor required to saturate 1 kilogram of dry air at various temperatures.

water vapor required for the saturation of 1 kilogram (2.2 pounds) of dry air at various temperatures is shown in Table 4–1. Note that for every 10°C (18°F) increase in temperature, the amount of water vapor needed for saturation almost doubles. Thus, roughly four times more water vapor is needed to saturate 30°C (86°F) air than 10°C (50°F) air.

The atmosphere behaves in much the same manner as our closed container. In nature, gravity, rather than a lid, prevents water vapor (and other gases) from escaping into space. Also like our container, water molecules are constantly evaporating from liquid surfaces (such as lakes or cloud droplets) and other vapor molecules are arriving. However, in nature a balance is not always achieved. More often than not, more water molecules are leaving the surface of a water puddle than are arriving, causing what meteorologists call *net evaporation*. By contrast, during the formation of fog, more water molecules are condensing than are evaporating from the tiny fog droplets, resulting in *net condensation*.

What determines whether the rate of evaporation exceeds the rate of condensation (net evaporation) or vice versa? One of the major factors is the temperature of the surface water, which in turn determines how much motion (kinetic energy) the water molecules possess. At higher temperatures the molecules have more energy and can more readily escape. Thus, under otherwise similar conditions, because hot water has more energy, it will evaporate faster than cold water.

TABLE 4-1 Saturation mixing ratio (at sea-level pressure)

Temperature °C (°F)	Saturation mixing ratio g/kg
−40 (−40)	0.1
−30 (−22)	0.3
−20 (−4)	0.75
−10 (14)	2
0 (32)	3.5
5 (41)	5
10 (50)	7
15 (59)	10
20 (68)	14
25 (77)	20
30 (86)	26.5
35 (95)	35
40 (104)	47

The other major factor determining which will dominate, evaporation or condensation, is the vapor pressure in the air around the liquid. Recall from our example of a closed container that vapor pressure determines the rate at which the water molecules return to the surface (condense). When the air is dry (low vapor pressure), the rate at which water molecules return to the liquid phase is low. However, when the air around a liquid has reached the saturation vapor pressure, the rate of condensation will be equal to the rate of evaporation. Thus, at saturation there is neither a net condensation nor a net evaporation. Therefore, all else being equal, net evaporation is greater when the air is dry (low vapor pressure) than when the air is humid (high vapor pressure).

In addition to vapor pressure and temperature, other factors exist in nature that affect rates of evaporation and condensation. Although these factors are of minimal importance to most of the processes operating at Earth's surface, they are significant in the atmosphere where clouds and precipitation form. Thus, we will revisit this idea when we consider the formation of clouds and precipitation in Chapter 5.

Students Sometimes Ask

Why do the sizes of snow piles seem to shrink a few days after a snowfall, even when the temperatures remain below freezing?

On clear, cold days following a snowfall, the air can be very dry. This fact, plus solar heating, causes the ice crystals to sublimate—turn from a solid to a gas. Thus, even without any appreciable melting, these accumulations of snow gradually get smaller.

Relative Humidity

GEODe ATMOSPHERE

Moisture and Cloud Formation

▸ Humidity: Water Vapor in the Air

The most familiar and, unfortunately, the most misunderstood term used to describe the moisture content of air is relative humidity. **Relative humidity** *is a ratio of the air's actual water-vapor content compared with the amount of water vapor required for saturation at that temperature (and pressure).* Thus, relative humidity indicates how near the air is to saturation rather than the actual quantity of water vapor in the air (see Box 4–2).

To illustrate, we see from Table 4–1 that at 25°C, air is saturated when it contains 20 grams of water vapor per kilogram of air. Thus, if the air contains 10 grams per kilogram on a 25°C day, the relative humidity is expressed as 10/20, or 50 percent. Further, if air with a temperature of 25°C had a water-vapor content of 20 grams per kilogram, the relative humidity would be expressed as 20/20 or 100 percent. On those occasions when the relative humidity reaches 100 percent, the air is said to be saturated.

How Relative Humidity Changes

Because relative humidity is based on the air's water-vapor content, as well as the amount of moisture required for saturation, it can be changed in either of two ways. First, relative humidity can be changed by the addition or removal of water vapor. Second, because the amount of moisture required for saturation is a function of air temperature, relative humidity varies with temperature. (Recall that the saturation vapor pressure is temperature dependent, such that at higher temperatures, it takes more water vapor to saturate air than at lower temperatures.)

Adding or Subtracting Moisture. Notice in Figure 4–10. that when water vapor is added to a parcel of air, its relative humidity increases until saturation occurs (100 percent relative humidity). What if even more moisture is added to this parcel of saturated air? Does the relative humidity exceed 100 percent? Normally, this situation does not occur. Instead, the excess water vapor condenses to form liquid water.

You may have experienced such a situation while taking a hot shower. The water leaving the shower is composed of very energetic (hot) molecules, which means that the rate of evaporation is high. As long as you run the shower, the process of evaporation continually adds water vapor to the unsaturated air in the bathroom. Therefore, if you stay in a hot shower long enough, the air eventually becomes saturated and the excess water vapor condenses on the mirror, window, tile, and other surfaces in the room.

In nature, moisture is added to the air mainly via evaporation from the oceans. However, plants, soil, and smaller bodies of water do make substantial contributions. Unlike your shower, however, the processes that add water vapor to the air generally do not operate at rates fast enough to cause saturation to occur directly. One exception is when you exhale on a cold winter day and "see your breath." What is happening is the warm moist air from your lungs mixes with the cold outside air, which has a very low saturation vapor pressure. Your breath has enough moisture to saturate a small quantity of cold outside air and the result is a miniature "cloud." Almost as fast as the "cloud" forms, it mixes with more of the dry outside air and quickly evaporates.

Changes with Temperature. The second condition that affects relative humidity is air temperature (see Box 4–3). Examine Figure 4–11 carefully. Note that in Figure 4–11a, when air at 20°C contains 7 grams of water vapor per kilogram, it has a relative humidity of 50 percent. This can be verified by referring to Table 4–1. Here we can see that at 20°C, air is saturated when it contains 14 grams of water vapor per kilogram of air. Because the air in Figure 4–11a contains 7 grams of water vapor, its relative humidity is 7/14 or 50 percent.

How does cooling affect relative humidity? When the flask in Figure 4–11a is cooled from 20° to 10°C, as shown in Figure 4–11b, the relative humidity increases from 50 to

BOX 4-2 Dry Air at 100 Percent Relative Humidity?

A common misconception relating to meteorology is the notion that air with a high relative humidity must have a greater water-vapor content than air with a lower relative humidity. Frequently, this is not the case (Figure 4–C). To illustrate, let us compare a typical January day at International Falls, Minnesota, to one in the desert near Phoenix, Arizona. On this hypothetical day the temperature in International Falls is a cold −10°C (14°F) and the relative humidity is 100 percent. By referring to Table 4–1, we can see that saturated −10°C (14°F) air has a water-vapor content (mixing ratio) of 2 grams per kilogram (g/kg). By contrast, the desert air at Phoenix on this January day is a warm 25°C (77°F), and the relative humidity is just 20 percent. A look at Table 4–1 reveals that 25°C (77°F) air has a saturation mixing ratio of 20 g/kg. Therefore, with a relative humidity of 20 percent, the air at Phoenix has a water-vapor content of 4 g/kg (20 grams × 20 percent). Consequently, the "dry" air at Phoenix actually contains twice the water vapor as the "wet" air at International Falls.

This should make clear why places that are very cold are also very dry. The low water-vapor content of frigid air (even when saturated) helps to explain why many arctic areas receive only meager amounts of precipitation and are sometimes referred to as "polar deserts." This also helps us understand why people frequently experience dry skin and chapped lips during the winter months. The water vapor content of cold air is low, even when compared to some hot, arid regions.

FIGURE 4-C Moisture content of hot air versus frigid air. Hot desert air with a low relative humidity generally has a higher water-vapor content than frigid air with a high relative humidity. *(Top photo by E. J. Tarbuck, bottom photo by Matt Duvall)*

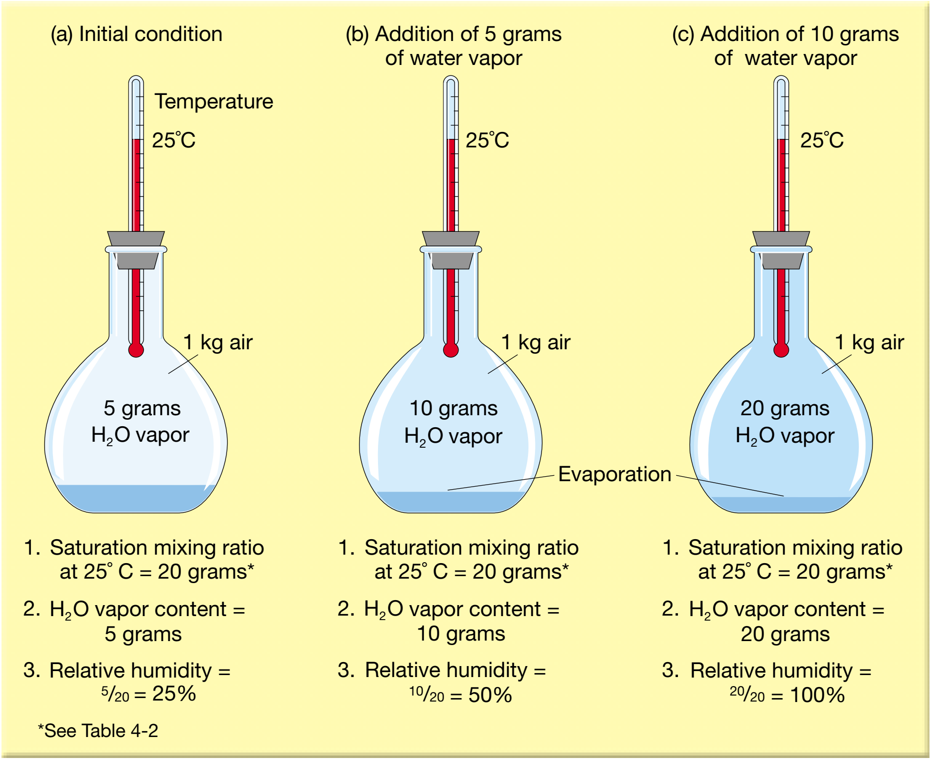


FIGURE 4-10 Relative humidity. At a constant temperature the relative humidity will increase as water vapor is added to the air. Here, the saturation mixing ratio remains constant at 20 grams per kilogram and the relative humidity rises from 25 to 100 percent as the water-vapor content increases.

100 percent. We can conclude from this that when the water-vapor content remains constant, a decrease in temperature results in an increase in relative humidity.

But there is no reason to assume that cooling would cease the moment the air reached saturation. What happens when the air is cooled below the temperature at which saturation occurs? Figure 4–11c illustrates this situation. Notice from Table 4–1 that when the flask is cooled to 0°C, the air is saturated at 3.5 grams of water vapor per kilogram of air. Because this flask originally contained 7 grams of water vapor, 3.5 grams of water vapor will condense to form liquid droplets that collect on the walls of the container. In the meantime, the relative humidity of the air inside remains at 100 percent. This raises an important concept. When air

BOX 4-3 Humidifiers and Dehumidifiers

In summer, stores sell *dehumidifiers.* As winter rolls around, these same merchants feature *humidifiers.* Why do you suppose so many homes are equipped with both a humidifier and a dehumidifier? The answer lies in the relationship between temperature and relative humidity. Recall that if the water-vapor content of air remains at a constant level, an increase in temperature lowers the relative humidity and a lowering of temperature increases the relative humidity.

During the summer months, warm, moist air frequently dominates the weather of the central and eastern United States. When hot and humid air enters a home, some of it circulates into the cool basement. As a result, the temperature of this air drops and its relative humidity increases. The result is a damp, musty-smelling basement. In response, the homeowner installs a dehumidifier to alleviate the problem. As air is drawn over the cold coils of the dehumidifier, water vapor condenses and collects in a bucket or flows down a drain. This process reduces the relative humidity and makes for a drier, more comfortable basement.

By contrast, during the winter months, outside air is cool and dry. When this air is drawn into the home, it is heated to room temperature. This process in turn causes the relative humidity to plunge, often to uncomfortably low levels of 40 percent or lower. Living with dry air can mean static electrical shocks, dry skin, sinus headaches, or even nose bleeds. Consequently, the homeowner may install a humidifier, which adds water to the air and increases the relative humidity to a more comfortable level.

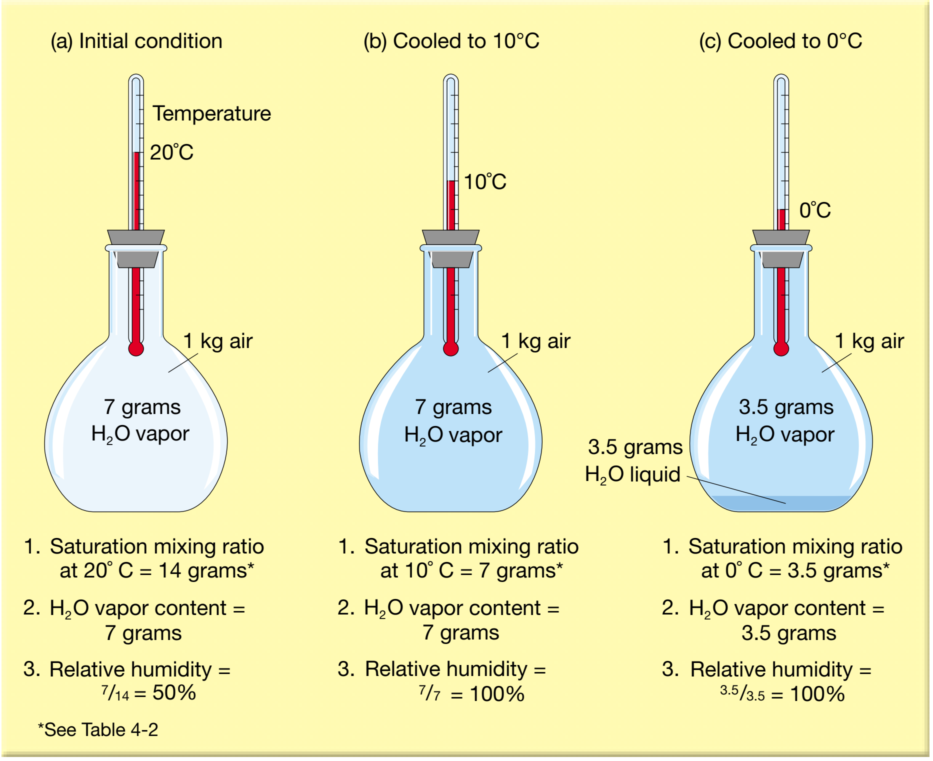


FIGURE 4-11 Relative humidity varies with temperature. When the water-vapor content (mixing ratio) remains constant, the relative humidity can be changed by increasing or decreasing the air temperature. In this example, when the temperature of the air in the flask was lowered from 20° to 10°C, the relative humidity increased from 50 to 100 percent. Further cooling (from 10° to 0°C) causes one-half of the water vapor to condense. In nature, cooling of air below its saturated mixing ratio generally causes condensation in the form of clouds, dew, or fog.

aloft is cooled below its saturation level, some of the water vapor condenses to form clouds. As clouds are made of liquid droplets, this moisture is no longer part of the water-vapor content of the air.

Let us return to Figure 4–11 and see what would happen should the flask in Figure 4–11a be heated to 35°C. From Table 4–1 we see that at 35°C, saturation occurs at 35 grams of water vapor per kilogram of air. Consequently, by heating the air from 20° to 35°C, the relative humidity would drop from 7/14 (50 percent) to 7/35 (20 percent).

We can summarize the effects of temperature on relative humidity as follows. When the water-vapor content of air remains at a constant level, a decrease in air temperature results in an increase in relative humidity, and an increase in temperature causes a decrease in relative humidity.

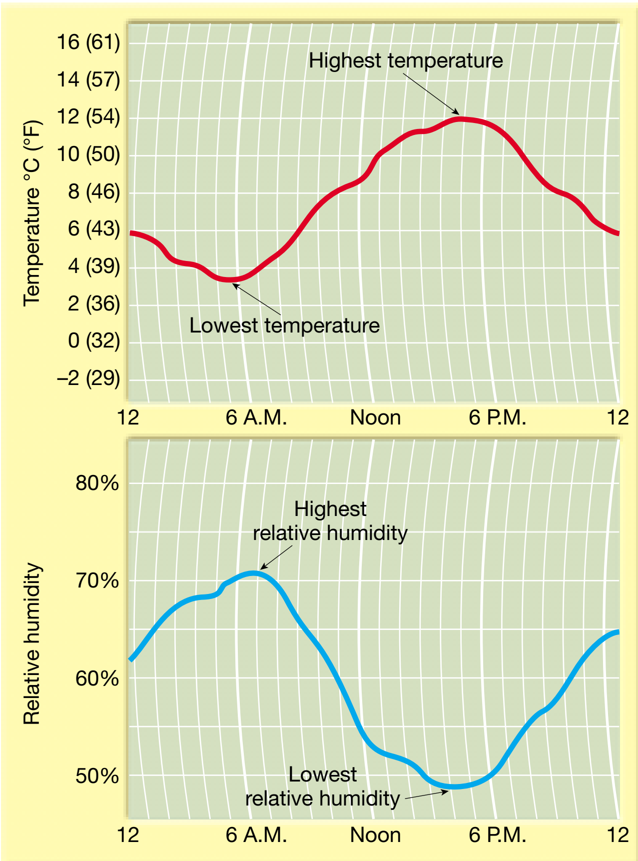
Natural Changes in Relative Humidity

In nature there are three major ways that air temperatures change (over relatively short time spans) to cause corresponding changes in relative humidity. These are:

1. Daily changes in temperatures (daylight versus nighttime temperatures).
2. Temperature changes that result as air moves horizontally from one location to another.
3. Temperature changes caused as air moves vertically in the atmosphere.

The importance of the last two processes in creating weather will be discussed later. The effect of the typical daily temperature cycle on relative humidity is shown in Figure 4–12. Notice that during the warmer midday period, relative

FIGURE 4-12 Typical daily variation in temperature and relative humidity during a spring day at Washington, DC.



humidity reaches its lowest level, whereas the cooler evening hours are associated with higher relative humidities. In this example, the actual water-vapor content (mixing ratio) of the air remains unchanged; only the relative humidity varies. Now we can better understand why a high relative humidity does not necessarily indicate a high water-vapor content.

Despite the previous example, we still describe air having a low relative humidity as being "dry" and vice versa. The use of the word "dry" in this context indicates that the air is far from being saturated. Thus, the rate of evaporation on a dry day is generally higher than on a humid day.

In summary, relative humidity indicates how near the air is to being saturated, whereas the air's mixing ratio denotes the actual quantity of water vapor contained in that air.

Students Sometimes Ask

Why do my lips get chapped in the winter?

During the winter months, outside air is comparatively cool and dry. When this air is drawn into the home, it is heated, which causes the relative humidity to plunge. Unless your home is equipped with a humidifier, you are likely to experience chapped lips and dry skin at that time of year.

Dew-Point Temperature

Moisture and Cloud Formation

▸ Humidity: Water Vapor in the Air

Another important measure of humidity is the dew-point temperature. The **dew point temperature,** or simply the **dew point,** is the temperature to which air needs to be cooled to reach saturation. For example, in Figure 4–11 the unsaturated air in the flask had to be cooled to 10°C before saturation occurred. Therefore, 10°C is the dew-point temperature for this air. In nature, cooling below the dew point causes water vapor to condense, typically as dew, fog, or clouds (Figure 4–13). The term *dew point* stems from the fact that during nighttime hours, objects near the ground often cool below the dew-point temperature and become coated with dew.*

Unlike relative humidity, which is a measure of how near the air is to being saturated, dew-point temperature is a measure of the *actual moisture content* of a parcel of air. Because the dew-point temperature is directly related to the amount of water vapor in the air, and because it is easy to determine, it is one of the most useful measures of humidity.

*Normally we associate dew with grass. Because of transpiration by the blades of grass, the relative humidity on a calm night is much higher near the grass than a few inches above the surface. Consequently, dew forms on grass before it does on most other objects.

FIGURE 4-13 Condensation, or "dew," occurs when a cold drinking glass chills the surrounding layer of air to the dew-point temperature. *(Photo by © Dorling Kindersley)*

Recall that the saturation vapor pressure is temperature-dependent and that for every 10°C (18°F) increase in temperature, the amount of water vapor needed for saturation doubles. Therefore, relatively cold *saturated air* (0°C or 32°F) contains about half the water vapor of *saturated air* having a temperature of 10°C (50°F) and roughly one-fourth that of hot *saturated air* with a temperature of 20°C (68°F). Because the dew point is the temperature at which saturation occurs, we can conclude that high dew-point temperatures equate to moist air and low dew-point temperatures indicate dry air (Table 4–2). More precisely, based on what we have learned about vapor pressure and saturation, we can state that for every 10°C (18°F) increase in the dew-point temperature, air contains about twice as much water vapor. Therefore, we know that when the dew-point temperature is 25°C (77°F), air contains about twice the water vapor as when the dew point is 15°C (59°F) and four times that of air with a dew point of 5°C (41°F).

Because the dew-point temperature is a good measure of the amount of water vapor in the air, it is the measure of atmospheric moisture that appears on a variety of weather

TABLE 4–2 Dew-point thresholds

Dew-point temperature	
≤ 10°F	Significant snowfall is inhibited.
≥ 55°F	Minimum for severe thunderstorms to form.
≥ 65°F	Considered humid by most people.
≥ 70°F	Typical of the rainy tropics.
≥ 75°F	Considered oppressive by most.

maps. Notice on the map in Figure 4–14 that many of the places located near the warm Gulf of Mexico have dew-point temperatures that exceed 70°F (21°C). When the dew point exceeds 65°F (18°C), it is considered humid by most people, and air with a dew point 75°F (24°C) or higher is oppressive. Also notice in Figure 4–14 that although the Southeast is dominated by humid conditions (dew points above 65°F), most of the remainder of the country is experiencing comparatively drier air.

Students Sometimes Ask

Is frost just frozen dew?

Contrary to popular belief, frost is not frozen dew. Rather, white frost (*hoar frost*) forms on occasions when saturation occurs at temperature of 0°C (32°F) or below (a temperature called the *frost point*). Thus, frost forms when water vapor changes directly from a gas into a solid (ice) without entering the liquid state. This process, called *deposition,* produces delicate patterns of ice crystals that often decorate windows during winter.

Humidity Measurement

Aside from the fact that humidity is important meteorologically, many of us are interested in humidity because it influences comfort (see Box 4–4). Here we will look at the various ways that humidity is measured.

Absolute humidity and mixing ratio are difficult to measure directly, but if the relative humidity is known, they can be readily computed from a table or graph. How is relative humidity measured?

A variety of instruments, called **hygrometers,** can be used to measure relative humidity. One of the simplest hygrometers, a **psychrometer,** consists of two identical thermometers mounted side by side (Figure 4–15). One thermometer, called the *wet bulb,* has a thin muslin wick tied around the end.

To use the psychrometer, the cloth wick is saturated with water and a continuous current of air is passed over the wick, either by swinging the instrument freely in the air or by fanning air past it. As a result, water evaporates from the wick, absorbing heat energy from the thermometer to do so, and the temperature of the wet bulb drops. The amount of cooling that takes place is directly proportional to the dryness of the air. The drier the air, the greater the cooling. Therefore, the larger the difference between the wet- and dry-bulb temperatures, the lower the relative humidity; the smaller the difference, the higher the relative humidity. If the air is saturated, no evaporation will occur, and the two thermometers will have identical readings.

Tables have been devised to obtain both the relative humidity and the dew-point temperature. (Please refer to Appendix C, Table C–1 and Table C–2.) All that is required is to record the air (dry-bulb) temperature and calculate the difference between the dry- and wet-bulb readings. The difference is known as the *depression of the wet bulb.* Assume, for instance, that the dry-bulb temperature is 20°C and that the wet-bulb reading after swinging or fanning is 14°C. To determine the relative humidity, find the dry-bulb temperature on the left-hand column of Table C–1 of Appendix C and the depression of the wet bulb across the top. The relative humidity is found where the two meet. In this example, the relative humidity is 51 percent. The dew point can be determined in the same way, using Table C–2. In this case, it would be 10°C.

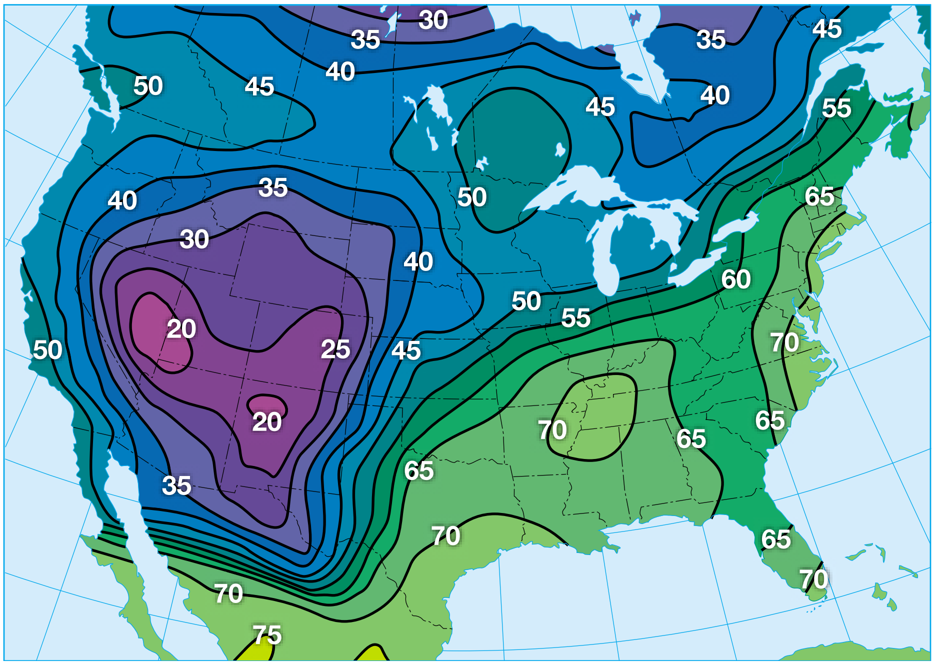


FIGURE 4-14 Surface map showing dew-point temperatures for September 15, 2005. Dew-point temperatures above 60°F dominate the southeastern United States, indicating that this region is blanketed with humid air.

BOX 4-4 Atmospheric Hazard: Humidity and Heat Stress

During a heat wave in 1995 more than 500 heat-related fatalities occurred in the greater Chicago area. Although this was an exceptional event, the stress of high summer temperatures and exposure to the Sun claim about 175 American lives in an average year.

High humidity contributes significantly to the discomfort people feel during a heat wave. Why are hot, muggy days so uncomfortable? Humans, like other mammals, are warm-blooded creatures who maintain a constant body temperature regardless of the temperature of the environment. One of the ways the body prevents overheating is by perspiring or sweating. However, this process does little to cool the body unless the perspiration can evaporate. It is the cooling created by the evaporation of perspiration that reduces body temperature. Because high humidity retards evaporation, people are more uncomfortable on a hot and humid day than on a hot and dry day.

Generally, temperature and humidity are the most important elements influencing summertime human comfort. Several indices combine these factors to establish the level or degree of comfort or discomfort. One index widely used by the National Weather Service was developed by R. G. Steadman and is called the *heat stress index*, or simply the *heat index*. It is a measure of *apparent temperature*, the air temperature that an individual perceives. It indicates how "hot" an average person feels given various combinations of temperature and relative humidity (Table 4–A).

For example, we can see from Table 4–A that if the air temperature is 90°F and the relative humidity is 60 percent, it should feel like 100°F. Note that as the relative humidity increases, the apparent temperature, and thus heat stress, increases as well. Further, when the relative humidity is low, the apparent temperature can have a value that is less than the actual air temperature.

To advise the public on the potential danger from heat stress, the National Weather Service uses the apparent temperature to determine the level of human discomfort, as shown in Table 4–A. This method categorizes the impact that various apparent temperatures will have on the well-being of individuals. It is important to note that factors such as the length of exposure to direct sunlight, the wind speed, and the general health of the individual greatly affect the amount of stress a person will experience. Further, while a period of hot, humid weather in New Orleans might be tolerable, a similar event in Minneapolis, Minnesota, would tax that population. This is because hot and humid weather is more taxing on people who live where it is less common than it is on people who live where prolonged periods of heat and humidity are the rule.

TABLE 4-A Heat index

Air Temperature (°F)	Relative Humidity (%) 40	45	50	55	60	65	70	75	80	85	90	95	100	With prolonged exposure and/or physical activity
110	136													**Extreme danger** Heat stroke or sunstroke highly likely
108	130	137												
106	124	130	137											
104	119	124	131	137										
102	114	119	124	130	137									**Danger** Sunstroke, muscle cramps, and/or heat exhaustion likely
100	109	114	118	124	129	136								
98	105	109	113	117	123	128	134							
96	101	104	108	112	116	121	126	132						
94	97	100	102	106	110	114	119	124	129	135				**Extreme caution** Sunstroke, muscle cramps, and/or heat exhaustion possible
92	94	96	99	101	105	108	112	116	121	126	131			
90	91	93	95	97	100	103	106	109	113	117	122	127	132	
88	88	89	91	93	95	98	100	103	106	110	113	117	121	
86	85	87	88	89	91	93	95	97	100	102	105	108	112	**Caution** Fatigue possible
84	83	84	85	86	88	89	90	92	94	96	98	100	103	
82	81	82	83	84	84	85	86	88	89	90	91	93	95	
80	80	80	81	81	82	82	83	84	84	85	86	86	87	

FIGURE 4-15 Sling psychrometer. This instrument is used to determine relative humidity and dew point. The dry-bulb thermometer gives the current air temperature. The thermometers are spun until the temperature of the wet-bulb thermometer (covered with a cloth wick) stops declining. Then both thermometers are read, and the data are used in conjunction with Tables C–1 and C–2 in Appendix C. *(Photo by E. J. Tarbuck)*

Another instrument used for measuring relative humidity, the *hair hygrometer,* can be read directly without using tables. The hair hygrometer operates on the principle that hair changes length in proportion to changes in relative humidity. Hair lengthens as relative humidity increases and shrinks as relative humidity drops. People with naturally curly hair experience this phenomenon, for in humid weather their hair lengthens and hence becomes curlier. The hair hygrometer uses a bundle of hairs linked mechanically to an indicator that is calibrated between 0 and 100 percent. Thus, we need only glance at the dial to read directly the relative humidity. Unfortunately, the hair hygrometer is less accurate than the psychrometer. Furthermore, it requires frequent calibration and is slow in responding to changes in humidity, especially at low temperatures.

A different type of hygrometer is used in remote-sensing instrument packages, such as radiosondes, that transmit upper-air observations back to ground stations. The electric hygrometer contains an electrical conductor coated with a moisture-absorbing chemical. It works on the principle that electric current flow varies as the relative humidity varies. Most surface weather stations have converted from traditional hygrometers to electric hygrometers of the type used for upper-air observations.

Students Sometimes Ask

What are the most humid cities in the United States?

As you might expect, the most humid cities in the United States are located near the ocean in regions that experience frequent onshore breezes. The record belongs to Quillayute, Washington, with an average relative humidity of 83 percent. However, many coastal cities in Oregon, Texas, Louisiana, and Florida also have average relative humidities that exceed 75 percent. Coastal cities in the Northeast tend to be somewhat less humid because they often experience air masses that originate over the drier, continental interior.

Adiabatic Temperature Changes

GEODe ATMOSPHERE **Moisture and Cloud Formation**

▶ The Basis of Cloud Formation: Adiabatic Cooling

So far we have considered some basic properties of water vapor and how its variability in the atmosphere is measured. We are now ready to examine the critical role that water vapor plays in our daily weather.

Recall that condensation occurs when water vapor is cooled enough to change to a liquid. Condensation may produce dew, fog, or clouds. Although each type of condensation is different, all require saturated air to form. As indicated earlier, saturation occurs either when sufficient water vapor is added to the air or, more commonly, when the air is cooled to its dew-point temperature.

Heat near Earth's surface is readily exchanged between the ground and the air above. As the ground loses heat in the evening (radiation cooling), dew may condense on the grass and fog may form in the air near the surface. Thus, surface cooling that occurs after sunset accounts for some condensation. However, cloud formation often takes place during the warmest part of the day. Clearly some other mechanism must operate aloft that cools air sufficiently to generate clouds.

The process that generates most clouds is easily visualized. Have you ever pumped up a bicycle tire with a hand pump and noticed that the pump barrel became very warm? When you applied energy to compress the air, the motion of the gas molecules increased and the temperature of the air rose. Conversely, if you allow air to escape from a bicycle tire, it expands; the gas molecules move less rapidly and the air cools. You have probably felt the cooling effect of the propellant gas expanding as you applied hair spray or spray deodorant.

The temperature changes just described, in which heat was neither added nor subtracted, are called **adiabatic temperature changes.** They result when air is compressed or allowed to expand. In summary, *when air is allowed to expand, it cools; when air is compressed, it warms*.

Adiabatic Cooling and Condensation

To simplify the following discussion, it helps if we imagine a volume of air enclosed in a thin elastic cover. Meteorologists call this imaginary volume of air a **parcel.** Typically, we consider a parcel to be a few hundred cubic meters in volume, and we assume that it acts independently of the surrounding air. It is also assumed that no heat is transferred into, or out of, the parcel. Although highly idealized, over short time spans, a parcel of air behaves in a manner much like an actual volume of air moving vertically in the atmosphere. In nature, sometimes the surrounding air infiltrates a vertically moving column of air, a process called **entrainment.** For the following discussion we will assume no mixing of this type is occurring.

Any time a parcel of air moves upward, it passes through regions of successively lower pressure. As a result, ascending air expands and it cools adiabatically. Unsaturated air cools at a constant rate of 10°C for every 1000 meters of ascent (5.5°F per 1000 feet). Conversely, descending air comes under increasing pressure and is compressed and heated 10°C for every 1000 meters of descent (Figure 4–16). This rate of cooling or heating applies only to vertically moving unsaturated air and is known as the **dry adiabatic rate** ("dry" because the air is unsaturated).

If a parcel of air rises high enough, it will eventually cool to its dew point. Here the process of condensation begins. The altitude at which a parcel reaches saturation and cloud formation begins is called the **lifting condensation level.** At the lifting condensation level an important thing happens: The latent heat that was absorbed by the water vapor when it evaporated is liberated. Although the parcel will continue to cool adiabatically, the release of this latent heat slows the rate of cooling. In other words, when a parcel of air ascends above the lifting condensation level, the rate of cooling is reduced because the release of latent heat partially offsets the cooling due to expansion. This slower rate of cooling caused by the release of latent heat is called the **wet adiabatic rate** of cooling ("wet" because the air is saturated).

Because the amount of latent heat released depends on the quantity of moisture present in the air (generally between 0 and 4 percent), the wet adiabatic rate varies from 5°C per 1000 meters for air with a high moisture content to 9°C per 1000 meters for air with a low moisture content. (Figure 4–17) illustrates the role of adiabatic cooling in the formation of clouds. Note that from the surface up to the lifting condensation level, the air cools at the faster, dry adiabatic rate. The slower, wet adiabatic rate commences at the point where condensation begins.

Processes That Lift Air

GEODe ATMOSPHERE **Moisture and Cloud Formation**
▸ Processes That Lift Air

To review, when air rises, it expands and cools adiabatically. If air is lifted sufficiently, it will eventually cool to its dew-point temperature, saturation will occur, and clouds will develop. But why does air rise on some occasions and not on others?

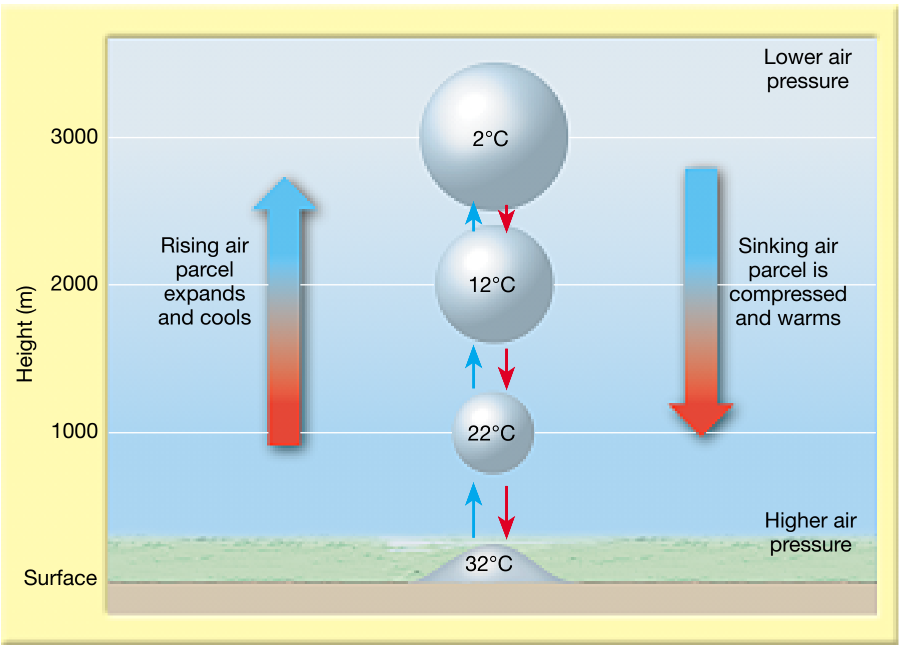


FIGURE 4-16 Whenever an unsaturated parcel of air is lifted it expands and cools at the *dry adiabatic rate* of 10°C per 1000 meters. Conversely, when air sinks, it is compressed and heats at the same rate.

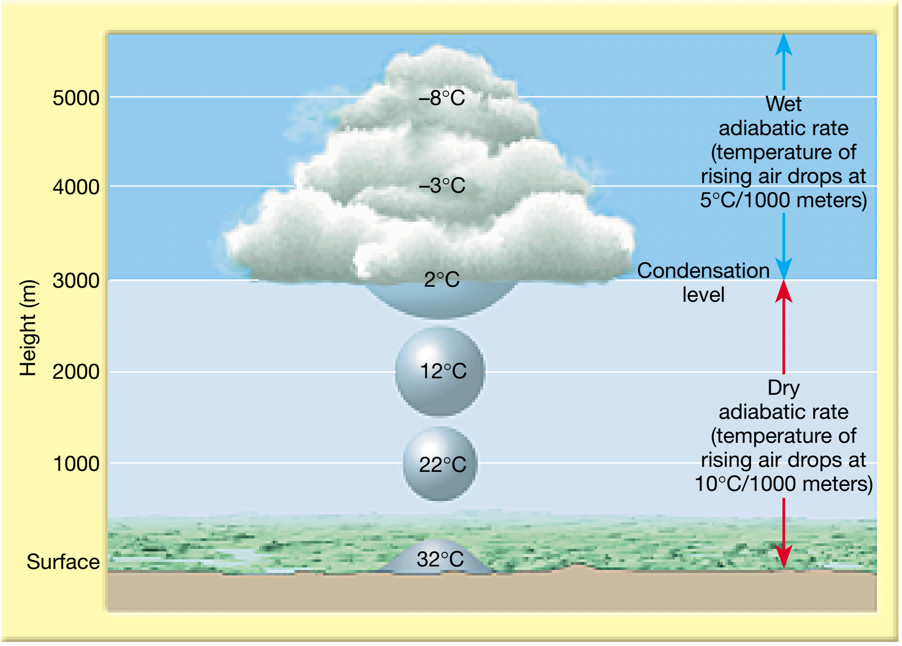


FIGURE 4-17 Rising air expands and cools at the dry adiabatic rate of 10°C per 1000 meters until the air reaches the dew point and condensation (cloud formation) begins. As air continues to rise, the latent heat released by condensation reduces the rate of cooling. The wet adiabatic rate is therefore always less than the dry adiabatic rate.

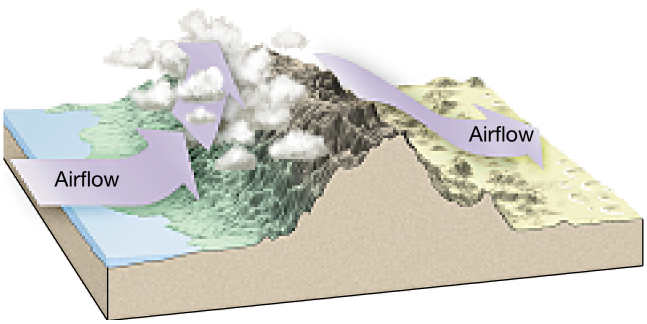
It turns out that in general the tendency is for air to resist vertical movement. Therefore, air located near the surface tends to stay near the surface, and air aloft tends to remain aloft. Exceptions to this rule, as we shall see, include conditions in the atmosphere that give air sufficient buoyancy to rise without the aid of outside forces. In many situations, however, when you see clouds forming, there is some mechanical phenomenon at work that forces the air to rise (at least initially).

We will look at four mechanisms that cause air to rise. These are:

1. *Orographic lifting*—air is forced to rise over a mountainous barrier.
2. *Frontal wedging*—warmer, less dense air, is forced over cooler, denser air.
3. *Convergence*—a pile-up of horizontal air flow results in upward movement.
4. *Localized convective lifting*—unequal surface heating causes localized pockets of air to rise because of their buoyancy.

In later chapters, we will consider other mechanisms that contribute to vertical airflow. In particular, horizontal airflow at upper levels significantly contributes to vertical lifting across the middle latitudes.

FIGURE 4-18 Orographic lifting occurs where air is forced over a topographic barrier.



Orographic Lifting

Orographic lifting occurs when elevated terrains, such as mountains, act as barriers to the flow of air (Figure 4–18). As air ascends a mountain slope, adiabatic cooling often generates clouds and copious precipitation. In fact, many of the rainiest places in the world are located on windward mountain slopes (see Box 4–5).

In addition to providing lift, mountains remove additional moisture in other ways. By slowing the horizontal flow of air, they cause convergence and retard the passage of storm systems. Moreover, the irregular topography of mountains enhances the differential heating that causes some localized convective lifting. These combined effects account for the generally higher precipitation associated with mountainous regions compared with surrounding lowlands.

By the time air reaches the leeward side of a mountain, much of its moisture has been lost. If the air descends, it warms adiabatically, making condensation and precipitation even less likely. As shown in Figure 4–18, the result can be a **rain shadow desert** (see Box 4–6). The Great Basin Desert of the western United States lies only a few hundred kilometers from the Pacific Ocean, but it is effectively cut off from the ocean's moisture by the imposing Sierra Nevada (Figure 4–19). The Gobi Desert of Mongolia, the Takla

BOX 4-5 Precipitation Records and Mountainous Terrain

Many of the rainiest places in the world are located on windward mountain slopes. Typically, these rainy areas occur because the mountains act as a barrier to the general circulation. Thus, the prevailing winds are forced to ascend the sloping terrain, thereby generating clouds and often abundant precipitation. A station at Mount Waialeale, Hawaii, for example, records the highest average annual rainfall in the world, some 1234 centimeters (486 inches). The station is located on the windward (northeast) coast of the island of Kauai at an elevation of 1569 meters (5148 feet). Incredibly, only 31 kilometers (19 miles) away lies sunny Barking Sands, with annual precipitation that averages less than 50 centimeters (20 inches).

The greatest recorded rainfall for a single 12-month period occurred at Cherrapunji, India, where an astounding 2647 centimeters (1042 inches), over 86 feet, fell. Cherrapunji, which is located at an elevation of 1293 meters (4309 feet), lies just north of the Bay of Bengal in an ideal location to receive the full effect of India's wet summer monsoon. Most of this rainfall occurred in the summer, particularly during the month of July, when a record 930 centimeters (366 inches) fell. For comparison, 10 times more rain fell in a single month at Cherrapunji, India, than falls in an average year at Chicago.

Because mountains can be sites of abundant precipitation, they are frequently very important sources of water. This is true for many dry locations in the western United States. Here the snow pack, which accumulates high in the mountains during the winter, is a major source of water for the summer season when precipitation is light and demand is great (Figure 4–D). Reservoirs in the Sierra Nevada, for example, accumulate and store spring runoff, which is then delivered to cities such as Los Angeles by way of an extensive network of canals. The record for greatest annual snowfall in the United States goes to the Mount Baker ski area north of Seattle, Washington, where 2896 centimeters (1140 inches) of snow were measured during the winter of 1998–1999.

FIGURE 4-D Heavy snow pack along Trail Ridge Road, in Colorado's Rocky Mountain National Park. *(Photo by Henry Lansford)*

Makan of China, and the Patagonia Desert of Argentina are other examples of deserts that exist because they are on the leeward sides of mountains.

Frontal Wedging

If orographic lifting were the only mechanism that forced air aloft, the relatively flat central portion of North America would be an expansive desert instead of the nation's breadbasket. Fortunately, this is not the case.

In central North America, masses of warm and cold air collide producing a **front.** Here the cooler, denser air acts as a barrier over which the warmer, less dense air rises. This process, called **frontal wedging,** is illustrated in (Figure 4–20).

It should be noted that weather-producing fronts are associated with storm systems called *middle-latitude cyclones*. Because these storms are responsible for producing a high percentage of the precipitation in the middle latitudes, we will examine them closely in Chapter 9.

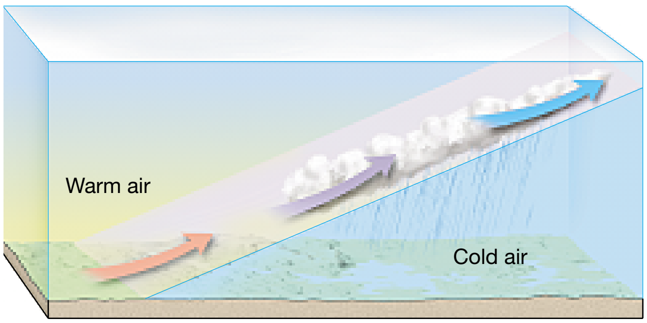
FIGURE 4-19 Rain-shadow desert. The arid conditions in California's Death Valley can be partially attributed to the adjacent mountains, which orographically remove the moisture from air originating over the Pacific. *(Photo by James E. Patterson/James Patterson Collection)*

Convergence

We saw that the collision of contrasting air masses forces air to rise. In a more general sense, whenever air in the lower troposphere flows together, lifting results. This phenomenon is called **convergence.** When air flows in from more than one direction, it must go somewhere. As it cannot go down, it goes up (Figure 4–21). This, of course, leads to adiabatic cooling and possibly cloud formation.

Convergence can also occur whenever an obstacle slows or restricts horizontal air flow (wind). We saw earlier that mountains slow winds and cause convergence. Further, when air moves from a relatively smooth surface, such as the ocean, onto an irregular landscape its speed is reduced. The result is a pileup of air (convergence). This is similar to what happens when people leave a well-attended sporting event and pileup results at the exits. When air converges, the air molecules do not simply squeeze closer together (like people); rather, there is a net upward flow.

FIGURE 4-20 Frontal wedging. Colder, denser air acts as a barrier over which warmer, less dense air rises.



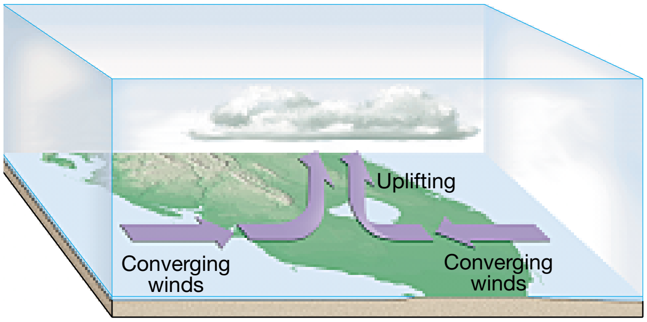


FIGURE 4-21 Convergence. When surface air converges, it increases in height to allow for the decreased area it occupies.

The Florida peninsula provides an excellent example of the role that convergence can play in initiating cloud development and precipitation. On warm days, the airflow is from the ocean to the land along both coasts of Florida. This leads to a pileup of air along the coasts and general convergence over the peninsula. This pattern of air movement and the uplift that results is aided by intense solar heating of the land. The result is that the peninsula of Florida experiences the greatest frequency of mid-afternoon thunderstorms in the United States (Figure 4–22).

More important, convergence as a mechanism of forceful lifting is a major contributor to the weather associated with middle-latitude cyclones and hurricanes. The low-level horizontal airflow associated with these systems is inward and upward around their centers. These important weather producers will be covered in more detail later, but for now remember that convergence near the surface results in a general upward flow.

FIGURE 4-22 Southern Florida viewed from the space shuttle. On warm days, airflow from the Atlantic Ocean and Gulf of Mexico onto the Florida peninsula generates many mid-afternoon thunderstorms. *(Photo by NASA/Media Services)*

BOX 4-6 Orographic Effects: Windward Precipitation and Leeward Rain Shadows

A good way to see the role that orographic lifting plays in the development of windward precipitation and leeward rain shadows is to examine a simplified hypothetical situation. In Figure 4–E, prevailing winds force warm, moist air over a 3000-meter-high mountain range. As the unsaturated air ascends the windward side of the range, it cools at the rate of 10°C per 1000 meters (dry adiabatic rate) until it reaches the dew-point temperature of 20°C. Because the dew-point temperature is reached at 1000 meters, we can say that this height represents the lifting condensation level and the height of the cloud base. Notice in Figure 4–E that above the lifting condensation level, latent heat is released, which results in a slower rate of cooling, called the *wet adiabatic rate.*

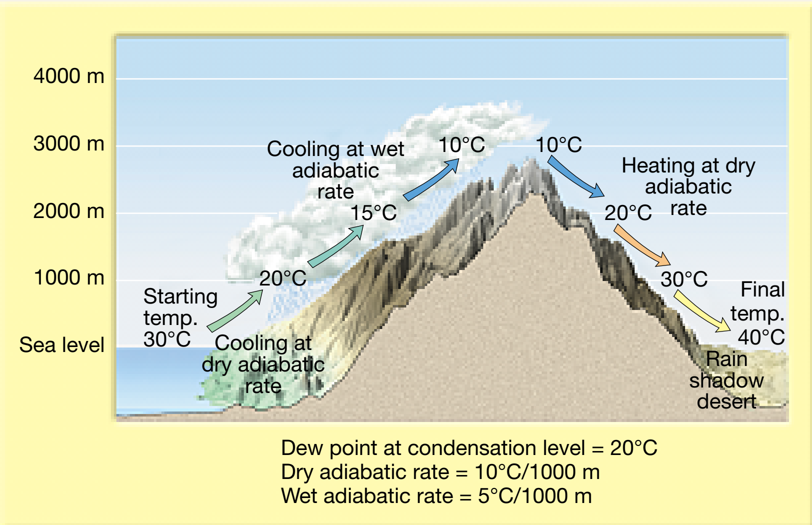
From the cloud base to the top of the mountain, water vapor within the rising air is condensing to form more and more cloud droplets. As a result, the windward side of the mountain range experiences abundant precipitation.

For simplicity, we will assume that the air that was forced to the top of the mountain is cooler than the surrounding air and hence begins to flow down the leeward slope of the mountain. As the air descends, it is compressed and *heated* at the dry adiabatic rate (Figure 4–E). Upon reaching the base of the mountain range, the temperature of the descending air has risen to 40°C, or 10°C warmer than the temperature at the base of the mountain on the windward side. The higher temperature on the leeward side is the result of the latent heat that was released during condensation as the air ascended the windward slope of the mountain range.

Two reasons account for the rain shadow commonly observed on leeward mountain slopes. First, water is extracted from air in the form of precipitation on the windward side. Second, the air on the leeward side is warmer than the air on the windward side. (Recall that whenever the temperature of air rises, the relative humidity drops.)

It should be emphasized that the hypothetical example provided here is only a rough approximation of conditions in the natural world. Most often, only a small percentage (sometimes none) of the moisture that condenses to form clouds actually falls as precipitation. Nevertheless, warmer and dryer conditions are the rule on the leeward side of a mountainous

FIGURE 4-E Orographic lifting and the formation of rain shadow deserts.



Localized Convective Lifting

On warm summer days, unequal heating of Earth's surface may cause pockets of air to be warmed more than the surrounding air (Figure 4–23). For instance, air above a paved parking lot will be warmed more than the air above an adjacent wooded park. Consequently, the parcel of air above the parking lot, which is warmer (less dense) than the surrounding air, will be buoyed upward. These rising parcels of warmer air are called *thermals.* Birds such as hawks and eagles use these thermals to carry them to great heights where they can gaze down on unsuspecting prey. People

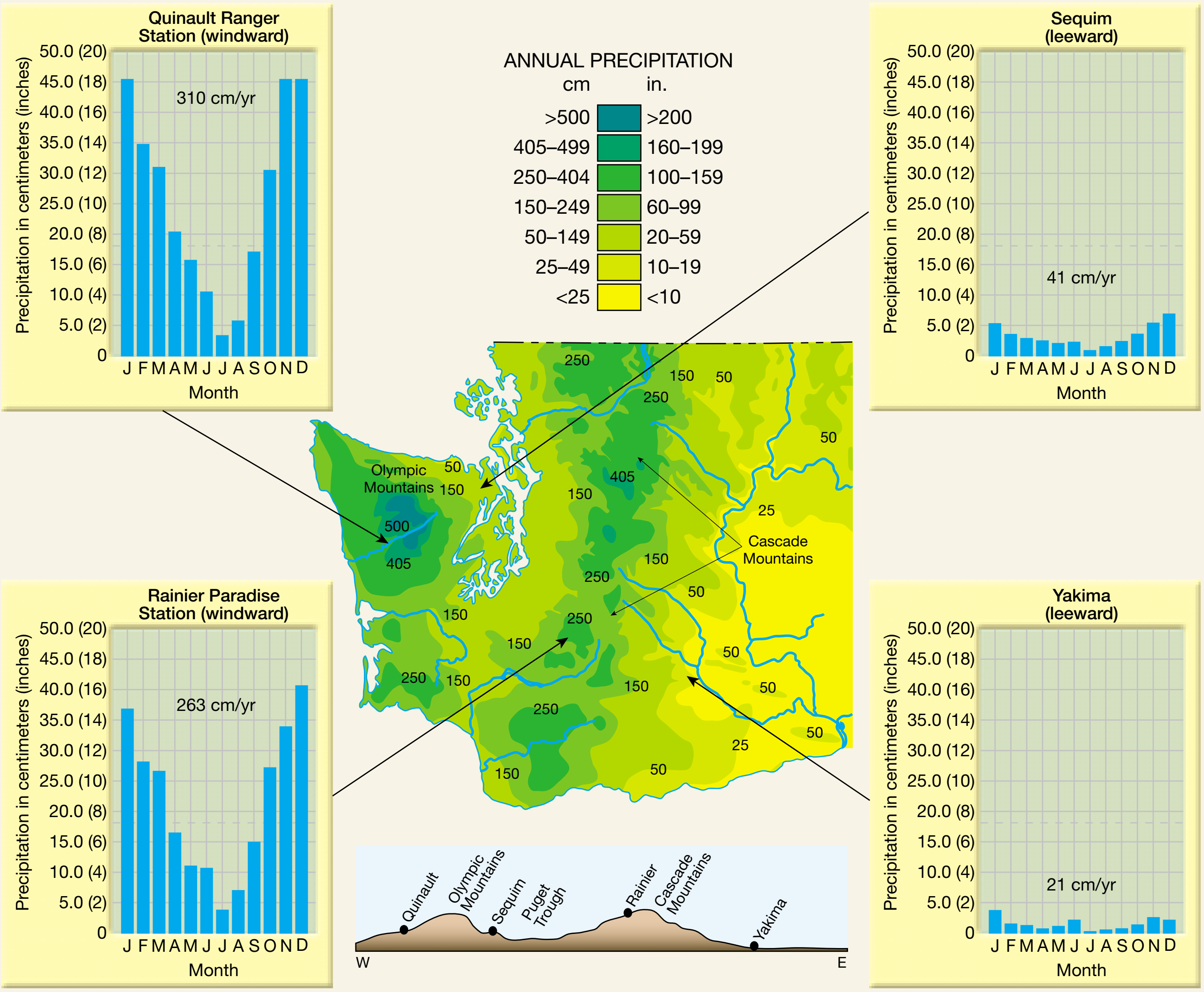


FIGURE 4-F Distribution of precipitation in western Washington State. Data from four stations provide examples of wetter windward locations and drier leeward rain shadows. (After Robert W. Christopherson).

barrier as compared to the rainier, windward side.

A classic example of windward precipitation and leeward rain shadows is found in western Washington State. As moist Pacific air flows inland over the Olympic and Cascade mountains, orographic precipitation is abundant (Figure 4–F). On the other hand, precipitation data for Sequim and Yakima indicate the presence of rain shadows on the leeward side of these highlands.

have learned to employ these rising parcels using hang gliders as a way to "fly."

The phenomenon that produces rising thermals is called **localized convective lifting.** When these warm parcels of air rise above the lifting condensation level, clouds form, which on occasion produce mid-afternoon rain showers. The height of clouds produced in this fashion is somewhat limited, for instability caused solely by unequal surface heating is confined to, at most, the first few kilometers of the atmosphere. Also, the accompanying rains, although occasionally heavy, are of short duration and widely scattered.

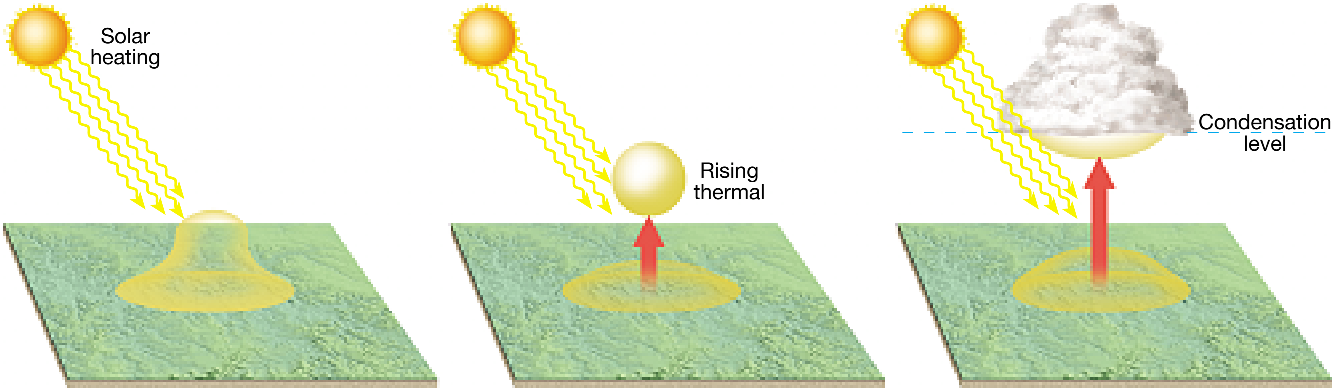


FIGURE 4-23 Localized convective lifting. Unequal heating of Earth's surface causes pockets of air to be warmed more than the surrounding air. These buoyant parcels of hot air rise, producing thermals, and if they reach the condensation level, clouds form.

Although localized convective lifting by itself is not a major producer of precipitation, the added buoyancy that results from surface heating contributes significantly to the lifting initiated by the other mechanisms. It should also be remembered that while the other mechanisms force air to rise, convective lifting occurs because the air is warmer (less dense) than the surrounding air and rises for the same reasons as a hot air balloon.

The Critical Weathermaker: Atmospheric Stability

GEODe ATMOSPHERE

Moisture and Cloud Formation

▶ The Critical Weathermaker: Atmospheric Stability

When air rises, it cools and eventually produces clouds. Why do clouds vary so much in size, and why does the resulting precipitation vary so much? The answers are closely related to the *stability* of the air.

Recall that a parcel of air can be thought of as having a thin flexible cover that allows it to expand but prevents it from mixing with the surrounding air (picture a hot-air balloon). If this parcel were forced to rise, its temperature would decrease because of expansion. By comparing the parcel's temperature to that of the surrounding air, we can determine its stability. If the parcel were *cooler* than the surrounding environment, it would be more dense; and if allowed to do so, it would sink to its original position. Air of this type, called **stable air,** resists vertical movement.

If, however, our imaginary rising parcel were *warmer* and hence less dense than the surrounding air, it would continue to rise until it reached an altitude where its temperature equaled that of its surroundings. This is exactly how a hot-air balloon works, rising as long as it is warmer and less dense than the surrounding air (Figure 4–24). This type of air is classified as **unstable air.** In summary, stability is a property of air that describes its tendency to remain in its original position (stable), or to rise (unstable).

Types of Stability

The stability of the atmosphere is determined by measuring the air temperature at various heights. Recall that this measure is called the *environmental lapse rate.* Do not confuse the environmental lapse rate with adiabatic temperature changes. The environmental lapse rate is the actual temperature of the atmosphere, as determined from observations made by radiosondes and aircraft. Adiabatic temperature

FIGURE 4-24 As long as air is warmer than its surroundings, it will rise. Hot-air balloons rise up through the atmosphere for this reason. *(Photo by Barbara Cushman Rowell/Mountain Light Photography, Inc.)*

changes are changes in temperature that a parcel of air would experience if it moved vertically through the atmosphere.

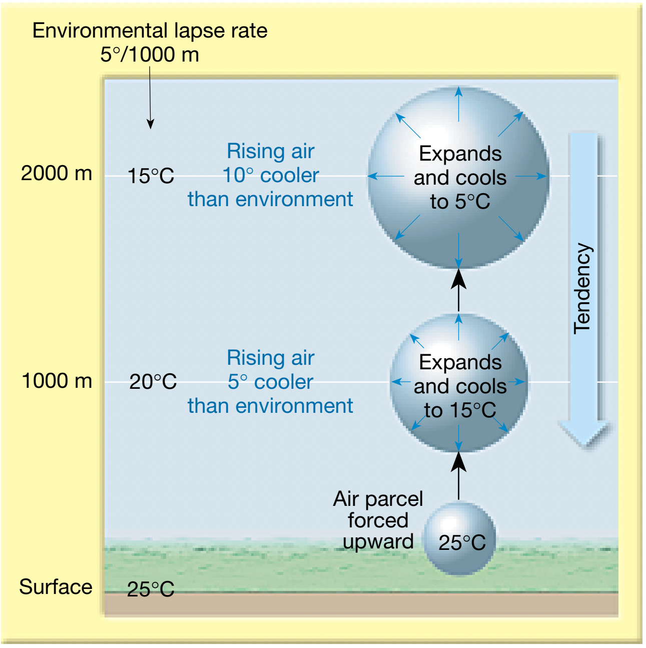
To illustrate how the stability of the atmosphere is determined, consider a situation in which the prevailing environmental lapse rate is 5°C per 1000 meters (Figure 4–25). Under this condition, when the air at the surface has a temperature of 25°C, the air at 1000 meters will be 5°C cooler, or 20°C, whereas the air at 2000 meters will have a temperature of 15°C, and so forth.

Examine Figure 4–25 and note that the air at the surface appears to be less dense than the air at 1000 meters, for it is 5°C warmer. However, if the air near the surface were to rise to 1000 meters, it would expand and cool at the dry adiabatic rate of 1°C per 100 meters. Therefore, on reaching 1000 meters, the temperature of the rising parcel would have dropped from 25°C to 15°C, a total of 10°C. Being 5°C cooler than its environment, it would be more dense and would tend to sink to its original position. Thus, we say that the air near the surface is *potentially cooler* than the air aloft and therefore it will not rise on its own.

By similar reasoning, if the air at 1000 meters subsided, adiabatic heating would increase its temperature 10°C by the time it reached the surface, making it warmer than the surrounding air; thus, its buoyancy would cause it to return to its original position. The air just described is stable and resists vertical movement.

With that background, we will now look at three fundamental conditions of the atmosphere: absolute stability, absolute instability, and conditional instability.

FIGURE 4-25 In a stable atmosphere, as an unsaturated parcel of air is lifted, it expands and cools at the dry adiabatic rate of 10°C per 1000 meters. Because the temperature of the rising parcel of air is lower than that of the surrounding environment, it will be heavier and, if allowed to do so, will sink to its original position.



Absolute Stability. Stated quantitatively, **absolute stability** prevails when *the environmental lapse rate is less than the wet adiabatic rate.* Figure 4–26 depicts this situation by using an environmental lapse rate of 5°C per 1000 meters and a wet adiabatic rate of 6°C per 1000 meters. Note that at 1000 meters the temperature of the surrounding air is 15°C and that the rising parcel of air has cooled by expansion to 10°C and so is the denser air. Even if this stable air were forced above the condensation level, it would remain cooler and denser than its environment and would have a tendency to return to the surface.

The most stable conditions occur when the temperature in a layer of air actually increases with altitude. When such a reversal occurs, a *temperature inversion* exists. Many circumstances can create temperature inversions. Temperature inversions frequently occur on clear nights as a result of radiation cooling at Earth's surface. Under these conditions an inversion is created because the ground and the air next to it will cool more rapidly than the air aloft.

Temperature inversions also occur in winter when warm air from the Gulf of Mexico invades the cold, snow-covered surface of the midcontinent. Anytime warmer air overlies cooler air, the resulting layer is extremely stable and resists appreciable vertical mixing. Because of this, temperature inversions are responsible for trapping pollutants in a narrow zone near Earth's surface. This idea will be explored more fully in Chapter 13.

Absolute Instability. At the other extreme, a layer of air is said to exhibit **absolute instability** when *the environmental lapse rate is greater than the dry adiabatic rate.* As shown in Figure 4–27, the ascending parcel of air is always warmer than its environment and will continue to rise because of its own buoyancy. Absolute instability occurs most often during the warmest months and on clear days when solar heating is intense. Under these conditions the lowermost layer of the atmosphere is heated to a much higher temperature than the air aloft. This results in a steep environmental lapse rate and a very unstable atmosphere.

Instability produced mainly by strong surface heating is generally confined to the first few kilometers of the atmosphere. Above this height the environmental lapse rate assumes a more "normal" value. Stated another way, the temperature drops more slowly with altitude, creating a more stable temperature regime. Consequently, clouds produced by surface heating lack great vertical height and thus rarely produce violent weather.

Conditional Instability. A more common type of atmospheric instability is called **conditional instability.** This situation prevails when *moist air has an environmental lapse rate between the dry and wet adiabatic rates* (between about 5° and 10°C per 1000 meters). Simply stated, the atmosphere is said to be conditionally unstable when it is *stable* with respect to an *unsaturated* parcel of air, but *unstable* with respect to a *saturated* parcel of air. Notice in

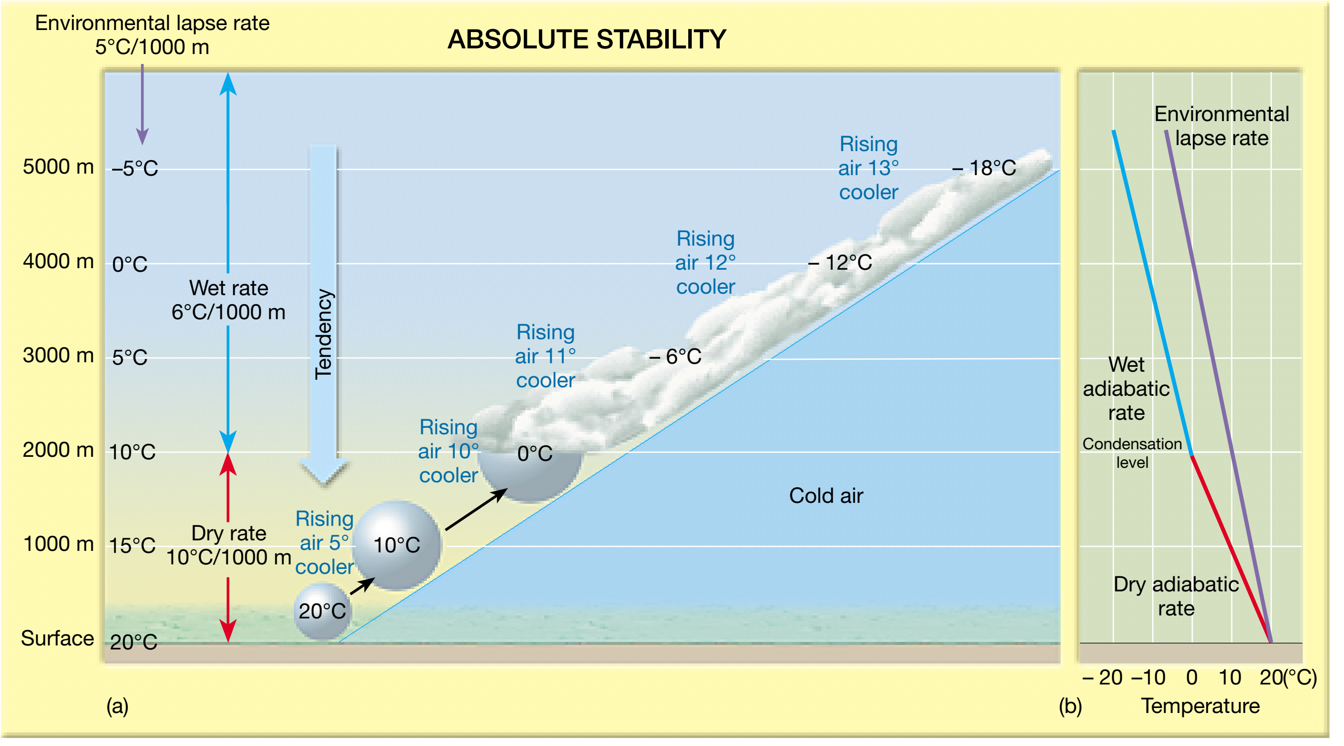


FIGURE 4-26 *Absolute stability* prevails when the environmental lapse rate is less than the wet adiabatic rate. (a) The rising parcel of air is always cooler and heavier than the surrounding air, producing stability. (b) Graphic representation of the conditions shown in part (a).

FIGURE 4-27 Illustration of *absolute instability* that develops when solar heating causes the lowermost layer of the atmosphere to be warmed to a higher temperature than the air aloft. The result is a steep environmental lapse rate that renders the atmosphere unstable. (b) Graphic representation of the conditions shown in part (a).

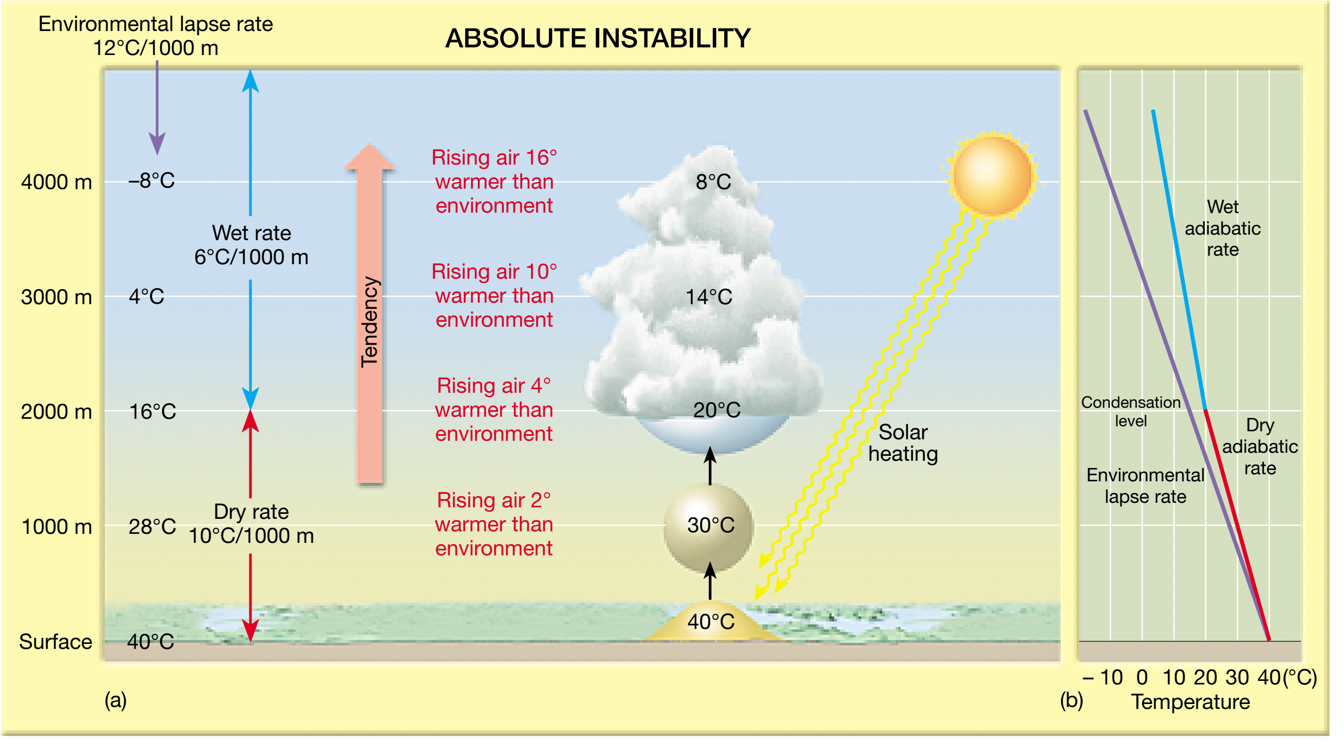


Figure 4–28 that the rising parcel of air is cooler than the surrounding air for the first 4000 meters. With the release of latent heat above the lifting condensation level, the parcel becomes warmer than the surrounding air. From this point along its ascent, the parcel will continue to rise without an outside force.

Thus, conditional instability depends on whether or not the rising air is saturated. The word "conditional" is used because the air must be forced upward before it reaches the level where it becomes unstable and rises on its own.

In summary, the stability of air is determined by measuring the temperature of the atmosphere at various heights (environmental lapse rate). In simple terms, a column of air is deemed unstable when the air near the bottom of this layer is significantly warmer (less dense) than the air aloft, indicating a steep environmental lapse rate. Under these conditions, the air actually turns over because the warm air below rises and displaces the colder air aloft. Conversely, the air is considered to be stable when the temperature decreases gradually with increasing altitude. The most stable conditions occur during a temperature inversion when the temperature actually increases with height. Under these conditions, there is very little vertical air movement.

Stability and Daily Weather

From the previous discussion, we can conclude that stable air resists vertical movement and that unstable air ascends freely because of its own buoyancy. How do these facts manifest themselves in our daily weather?

When stable air is forced aloft, the clouds that form are widespread and have little vertical thickness compared with their horizontal dimension. Precipitation, if any, is light to moderate. In contrast, clouds associated with unstable air are towering and are usually accompanied by heavy precipitation. Thus, we can conclude that on a dreary and overcast day with light drizzle, stable air was forced aloft. Conversely, on a day when cauliflower-shaped clouds appear to be growing

FIGURE 4-28 Illustration of *conditional instability,* where warm air is forced to rise along a frontal boundary. Note that the environmental lapse rate of 9°C per 1000 meters lies between the dry and wet adiabatic rates. (a) The parcel of air is cooler than the surrounding air up to nearly 3000 meters, where its tendency is to sink toward the surface (stable). Above this level, however, the parcel is warmer than its environment and will rise because of its own buoyancy (unstable). Thus, when conditionally unstable air is forced to rise, the result can be towering cumulus clouds. (b) Graphic representation of the conditions shown in part (a).

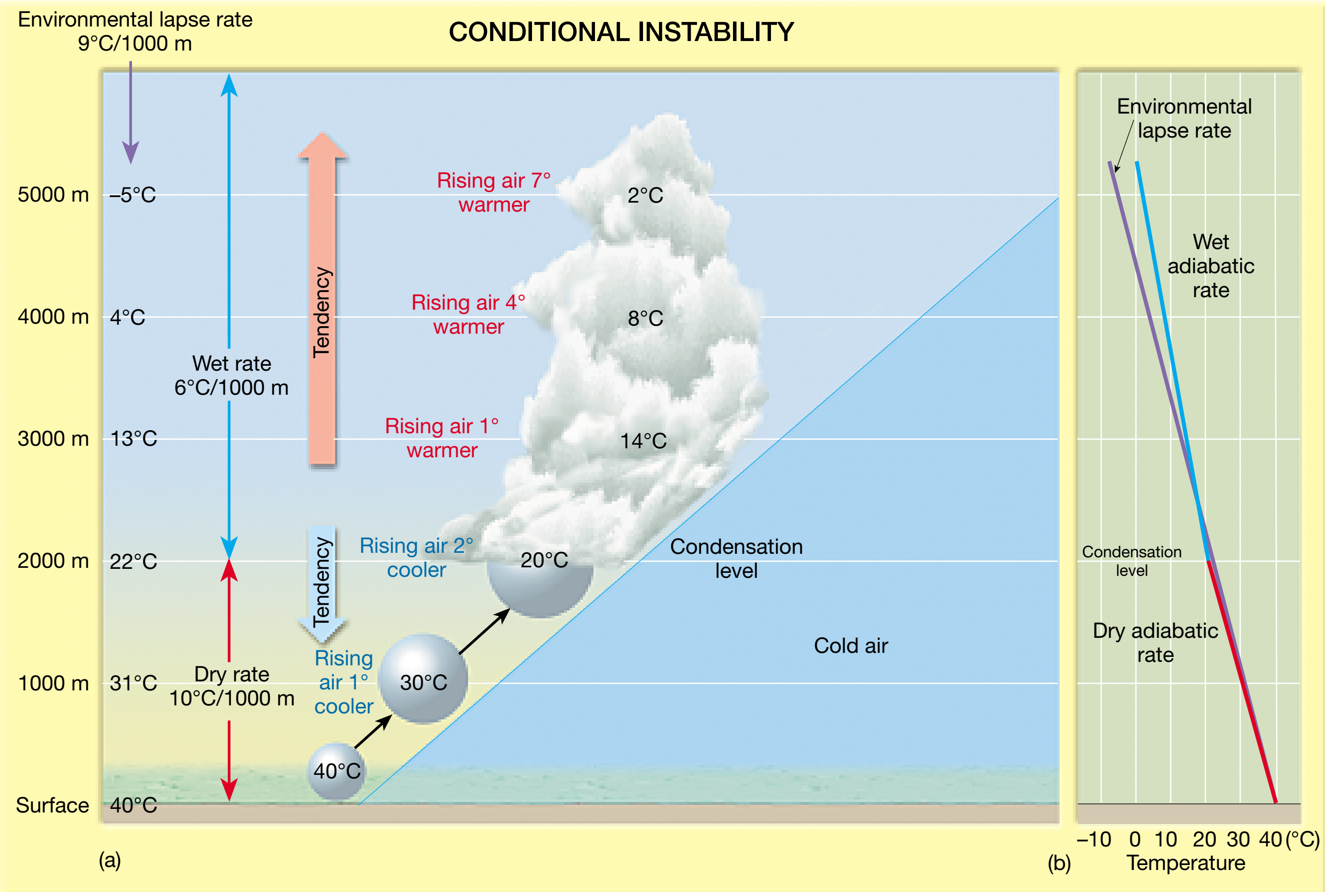


FIGURE 4-29 Cauliflower-shaped clouds provide evidence of unstable conditions in the atmosphere. *(Photo by Tom Bean/DRK Photo)*

as if bubbles of hot air were surging upward, we can be relatively certain that the atmosphere is unstable (Figure 4–29).

As noted earlier, the most stable conditions occur during a temperature inversion when temperature increases with height. In this situation, the air near the surface is cooler and heavier than the air aloft, and therefore little vertical mixing occurs between the layers. Because pollutants are generally added to the air from below, a temperature inversion confines them to the lowermost layer, where their concentration will continue to increase until the temperature inversion dissipates. Widespread fog is another sign of stability. If the layer containing fog were mixing freely with the "dry" layer above, evaporation would quickly eliminate the foggy condition.

How Stability Changes

Recall that the higher (steeper) the environmental lapse rate, the more rapidly the temperature drops with increasing altitude. Therefore, any factor that causes air near the surface to become warmed in relation to the air aloft increases instability. The opposite is also true; any factor that causes the surface air to be chilled results in the air becoming more stable. Also recall that the most stable conditions occur when the temperature in a layer of air actually increases with altitude.

Instability is enhanced by the following:

1. Intense solar heating warming the lowermost layer of the atmosphere.
2. The heating of an air mass from below as it passes over a warm surface.
3. General upward movement of air caused by processes such as orographic lifting, frontal wedging, and convergence.
4. Radiation cooling from cloud tops.

Stability is enhanced by the following:

1. Radiation cooling of Earth's surface after sunset.
2. The cooling of an air mass from below as it traverses a cold surface.
3. General subsidence within an air column.

Note that most processes that alter stability result from temperature changes caused by horizontal or vertical air movement, although daily temperature changes are important too. In general, any factor that increases the environmental lapse rate renders the air more unstable, whereas any factor that reduces the environmental lapse rate increases the air's stability.

Temperature Changes and Stability

As stated earlier, on a clear day when there is abundant surface heating, the lower atmosphere often becomes warmed sufficiently to cause parcels of air to rise. After the Sun sets, surface cooling generally renders the air stable again.

Similar changes in stability occur as air moves horizontally over a surface having markedly different temperatures. In the winter, warm air from the Gulf of Mexico moves northward over the cold, snow-covered Midwest. Because the air is cooled from below, it becomes more stable, often producing widespread fog.

The opposite occurs when wintertime polar air moves southward over the open waters of the Great Lakes. Although we would die of hypothermia in a few minutes if we fell into these cold waters (perhaps 5°C or 41°F), compared to the frigid air temperatures, these waters are warm. Recall from our discussion of vapor pressure that the temperature of the water surface and the vapor pressure of the surrounding air determine the rate at which water will evaporate. Because the water of the Great Lakes is comparatively warm and because the polar air is dry (low vapor pressure) the rate of evaporation will be high. The moisture and heat added to the frigid polar air from the water below are enough to make it unstable and generate the clouds that produce heavy snowfalls on the downwind shores of these lakes (called "lake-effect snows"—see Chapter 8).

Radiation Cooling from Clouds. On a smaller scale the loss of heat by radiation from cloud tops during evening hours adds to their instability and growth. Unlike air, which is a poor radiator of heat, cloud droplets emit energy to space nearly as well as does Earth's surface. Towering clouds that owe their growth to surface heating lose that source of energy at sunset. After sunset, however, radiation cooling at their tops steepens the lapse rate near the top of the cloud and can lead to additional upward flow of warmer parcels from below. This process is believed responsible for producing nocturnal thunderstorms from clouds whose growth prematurely ceased at sunset.

Vertical Air Movement and Stability

Vertical movements of air also influence stability. When there is a general downward airflow, called **subsidence,** the upper portion of the subsiding layer is heated by compression, more so than the lower portion. Usually the air near the surface is not involved in the subsidence and so its temperature remains unchanged. The net effect is to stabilize the air, for the air aloft is warmed in relation to the surface air. The warming effect of a few hundred meters of subsidence is enough to evaporate the clouds found in any layer of the atmosphere. Thus, one sign of subsiding air is a deep blue, cloudless sky. Subsidence can also produce a temperature inversion aloft. The most intense and prolonged temperature inversions and associated air pollution episodes are caused by subsidence, a topic discussed more fully in Chapter 13.

Upward movement of air generally enhances instability, particularly when the lower portion of the rising layer has a higher moisture content than the upper portion, which is usually the situation. As the air moves upward, the lower portion becomes saturated first and cools at the lesser wet adiabatic rate. The net effect is to increase the lapse rate within the rising layer. This process is especially important in producing the instability associated with thunderstorms. In addition, recall that conditionally unstable air can become unstable if lifted sufficiently.

In summary, the role of stability in determining our daily weather cannot be overemphasized. The air's stability, or lack of it, determines to a large degree whether clouds develop and produce precipitation and whether that precipitation will come as a gentle shower or a violent downpour. In general, when stable air is forced aloft, the associated clouds have little vertical thickness, and precipitation, if any, is light. In contrast, clouds associated with unstable air are towering and are frequently accompanied by heavy precipitation.

Chapter Summary

- The unending circulation of Earth's water supply is called the *hydrologic cycle* (or water cycle). The cycle illustrates the continuous movement of water from the oceans to the atmosphere, from the atmosphere to the land, and from the land back to the sea.
- *Water vapor,* an odorless, colorless gas, can change from one state of matter (solid, liquid or gas) to another at the temperatures and pressures experienced on Earth. The heat energy involved in the change of state of water is often measured in *calories.* The processes involved in changes of state include *evaporation* (liquid to gas), *condensation* (gas to liquid), *melting* (solid to liquid), *freezing* (liquid to solid), *sublimation* (solid to gas), and *deposition* (gas to solid). During each change, *latent* (hidden, or stored) *heat* energy is either absorbed or released.
- *Humidity* is the general term used to describe the amount of water vapor in the air. The methods used to express humidity quantitatively include (1) *absolute humidity,* the mass of water vapor in a given volume of air, (2) *mixing ratio,* the mass of water vapor in a unit of

air compared to the remaining mass of dry air, (3) *vapor pressure,* that part of the total atmospheric pressure attributable to its water-vapor content, (4) *relative humidity,* the ratio of the air's actual water-vapor content compared with the amount of water vapor required for saturation at that temperature, and (5) *dew point,* the temperature to which a parcel of air would need to be cooled to reach saturation. When air is *saturated,* the pressure exerted by the water vapor, called the *saturation vapor pressure,* produces a balance between the number of water molecules leaving the surface of the water and the number returning. Because the saturation vapor pressure is temperature-dependent, at higher temperatures more water vapor is required for saturation to occur.

- Relative humidity can be changed in two ways: (1) by changing the amount of moisture in the air or (2) by changing the air's temperature. Adding moisture to the air while keeping the temperature constant increases the relative humidity. Removing moisture lowers the relative humidity. When the water vapor content of air remains at a constant level, a decrease in air temperature results in an increase in relative humidity, and an increase in temperature causes a decrease in relative humidity. In nature there are three major ways that air temperatures change to cause corresponding changes in relative humidity: (1) daily (daylight versus nighttime) changes in temperature, (2) temperature changes that result as air moves horizontally from one location to another, and (3) changes caused as air moves vertically in the atmosphere.
- An important concept related to relative humidity is the *dew-point temperature* (or simply *dew point*), which is the temperature to which a parcel of air would need to be cooled to reach saturation. Unlike relative humidity, which is a measure of how near the air is to being saturated, dew-point temperature is a measure of the air's actual moisture content. High dew-point temperatures equate to moist air, and low dew-point temperatures indicate dry air. Because the dew-point temperature is a good measure of the amount of water vapor in the air, it is the measure of atmospheric moisture that appears on daily weather maps.
- A variety of instruments, called *hygrometers,* can be used to measure relative humidity. One of the simplest hygrometers, a *psychrometer,* consists of two identical thermometers mounted side by side. One thermometer, called the wet-bulb thermometer, has a thin muslin wick tied around the bulb. After spinning or fanning air past the instrument and noting the difference between the dry- and wet-bulb readings (known as the depression of the wet bulb), tables are consulted to determine the relative humidity. A second instrument, the *hair hygrometer,* can be read directly without using tables.
- When air is allowed to expand, it cools. When air is compressed, it warms. Temperature changes produced in this manner, in which heat is neither added nor subtracted, are called *adiabatic temperature changes.* The rate of cooling or warming of vertically moving unsaturated ("dry") air is 10°C for every 1000 meters (5.5°F per 1000 feet), the *dry adiabatic rate.* At the *lifting condensation level* (the altitude where a rising parcel of air has reached saturation and cloud formation begins), latent heat is released, and the rate of cooling is reduced. The slower rate of cooling, called the *wet adiabatic rate* of cooling ("wet" because the air is saturated) varies from 5°C per 1000 meters for air, with a high-moisture content to 9°C per 1000 meters for air with a low-moisture content.
- When air rises, it expands and cools adiabatically. If air is lifted sufficiently high, it will eventually cool to its dew-point temperature, and clouds will develop. Four mechanisms that cause air to rise are (1) *orographic lifting,* where air is forced to rise over a mountainous barrier, (2) *Frontal wedging,* where warmer, less dense air is forced over cooler, denser air along a *front,* (3) *convergence,* a pile-up of horizontal airflow resulting in an upward flow, and (4) *localized convective lifting,* where unequal surface heating causes localized pockets of air to rise because of their buoyancy.
- When air rises, it cools and can eventually produce clouds. *Stable air* resists vertical movement, whereas *unstable air* rises because of its buoyancy. The stability of air is determined by knowing the *environmental lapse rate,* the temperature of the atmosphere at various heights. The three fundamental conditions of the atmosphere are (1) *absolute stability,* when the environmental lapse rate is less than the wet adiabatic rate, (2) *absolute instability,* when the environmental lapse rate is greater than the dry adiabatic rate, and (3) *conditional instability,* when moist air has an environmental lapse rate between the dry and wet adiabatic rates. In general, when stable air is forced aloft, the associated clouds have little vertical thickness, and precipitation, if any, is light. In contrast, clouds associated with unstable air are towering and frequently accompanied by heavy rain.
- Any factor that causes air near the surface to become warmed in relation to the air aloft increases the air's instability. The opposite is also true; any factor that causes the surface air to be chilled results in the air becoming more stable. Most processes that alter atmospheric stability result from temperature changes caused by horizontal or vertical air movements, although daily temperature changes are important too. Changes in stability occur as air moves horizontally over a surface having a markedly different temperature than the air. Furthermore, *subsidence* (a general downward airflow) generally stabilizes the air, while upward air movement enhances instability.

Vocabulary Review

absolute humidity (p. 103)
absolute instability (p. 121)
absolute stability (p. 121)
adiabatic temperature changes (p. 114)
calorie (p. 101)
condensation (p. 102)
conditional instability (p. 121)
convergence (p. 117)
deposition (p. 102)
dew point (p. 110)
dew point temperature (p. 110)
dry adiabatic rate (p. 114)
entrainment (p. 114)
evaporation (p. 102)
front (p. 116)
frontal wedging (p. 116)
humidity (p. 103)
hydrologic cycle (p. 98)
hygrometer (p. 111)
latent heat (p. 101)
latent heat of condensation (p. 102)
latent heat of vaporization (p. 102)
lifting condensation level (p. 114)
localized convective lifting (p. 119)
mixing ratio (p. 104)
orographic lifting (p. 115)
parcel (p. 114)
psychrometer (p. 111)
rain shadow desert (p. 115)
relative humidity (p. 106)
saturation (p. 104)
saturation vapor pressure (p. 104)
stable air (p. 120)
sublimation (p. 102)
subsidence (p. 125)
transpiration (p. 98)
unstable air (p. 120)
vapor pressure (p. 104)
wet adiabatic rate (p. 114)

Review Questions

1. Describe the movement of water through the hydrologic cycle.
2. The quantity of water lost to evaporation over the oceans is not equaled by precipitation. Why, then, does the sea level not drop?
3. Summarize the processes by which water changes from one state to another. Indicate whether heat is absorbed or liberated.
4. After reviewing Table 4–1, write a generalization relating temperature and the amount of water vapor needed to saturate the air.
5. How do absolute humidity and mixing ratio differ? What do they have in common? How is relative humidity different from absolute humidity and the mixing ratio?
6. Refer to Figure 4–12 and then answer the following questions.
   a. When is relative humidity highest during a typical day? When is it lowest?
   b. At what time of day would dew most likely form?
   c. Write a generalization relating changes in air temperature to changes in relative humidity.
7. If temperature remains unchanged, and if the mixing ratio decreases, how will relative humidity change?
8. How much more water vapor does a mass of air having a dew-point temperature of 24°C (75°F) contain than air having a dew-point temperature of 4°C (39°F)?
9. Explain the principle of both the psychrometer and the hair hygrometer.
10. What are the disadvantages of the hair hygrometer? Does this instrument have any advantages over the psychrometer?
11. What name is given to the processes whereby the temperature of the air changes without the addition or subtraction of heat?
12. At what rate does unsaturated air cool when it rises through the atmosphere?
13. Why does air expand as it moves upward through the atmosphere?
14. Explain why air warms whenever it sinks.
15. Why does the adiabatic rate of cooling change when condensation begins? Why is the wet adiabatic rate not a constant figure?
16. The contents of an aerosol can are under very high pressure. When you push the nozzle on such a can, the spray feels cold. Explain.
17. How do orographic lifting and frontal wedging act to force air to rise?
18. Explain why the Great Basin area of the western United States is dry. What term is applied to such a situation?
19. How is localized convective lifting different from the other three processes that cause air to rise?
20. How does stable air differ from unstable air?
21. Explain the difference between the environmental lapse rate and adiabatic cooling.
22. How is the stability of air determined?
23. Write a statement relating the environmental lapse rate to stability.
24. What weather conditions would lead you to believe that air is unstable?
25. List four ways instability can be enhanced.
26. List three ways stability can be enhanced.

Problems

1. Using Table 4–1, answer the following:
   - **a.** If a parcel of air at 25°C contains 10 grams of water vapor per kilogram of air, what is its relative humidity?
   - **b.** If a parcel of air at 35°C contains 5 grams of water vapor per kilogram of air, what is its relative humidity?
   - **c.** If a parcel of air at 15°C contains 5 grams of water vapor per kilogram of air, what is its relative humidity?
   - **d.** If the temperature of this same parcel of air dropped to 5°C, how would its relative humidity change?
   - **e.** If 20°C air contains 7 grams of water vapor per kilogram of air, what is its dew point?
2. Using the standard tables (Appendix C Tables C–1 and C–2), determine the relative humidity and dew-point temperature if the dry-bulb thermometer reads 22°C and the wet-bulb thermometer reads 16°C. How would the relative humidity and dew point change if the wet-bulb reading were 19°C?
3. If unsaturated air at 20°C were to rise, what would its temperature be at a height of 500 meters? If the dew-point temperature at the lifting condensation level were 11°C, at what elevation would clouds begin to form?
4. Using Figure 4–30, answer the following. (*Hint:* Read Box 4–6.)
   - **a.** What is the elevation of the cloud base?
   - **b.** What is the temperature of the ascending air when it reaches the top of the mountain?
   - **c.** What is the dew-point temperature of the rising air at the top of the mountain? (Assume 100 percent relative humidity.)
   - **d.** Estimate the amount of water vapor that must have condensed (in grams per kilogram) as the air moved from the cloud base to the top of the mountain.
   - **e.** What will the temperature of the air be if it descends to point B? (Assume that the moisture that condensed fell as precipitation on the windward side of the mountain.)
   - **f.** What is the approximate capacity of the air to hold water vapor at point B?
   - **g.** Assuming that no moisture was added or subtracted from the air as it traveled downslope, estimate the relative humidity at point B.
   - **h.** What is the *approximate* relative humidity at point A? (Use the dew-point temperature at the condensation level for the surface dew point.)
   - **i.** Give *two* reasons for the difference in relative humidity between points A and B.
   - **j.** Needles, California, is situated on the dry leeward side of a mountain range similar to the position of point B. What term describes this situation?
5. If a lake is ice covered, what is the temperature of the water at the bottom of the lake? (*Hint:* See Box 4–1.)

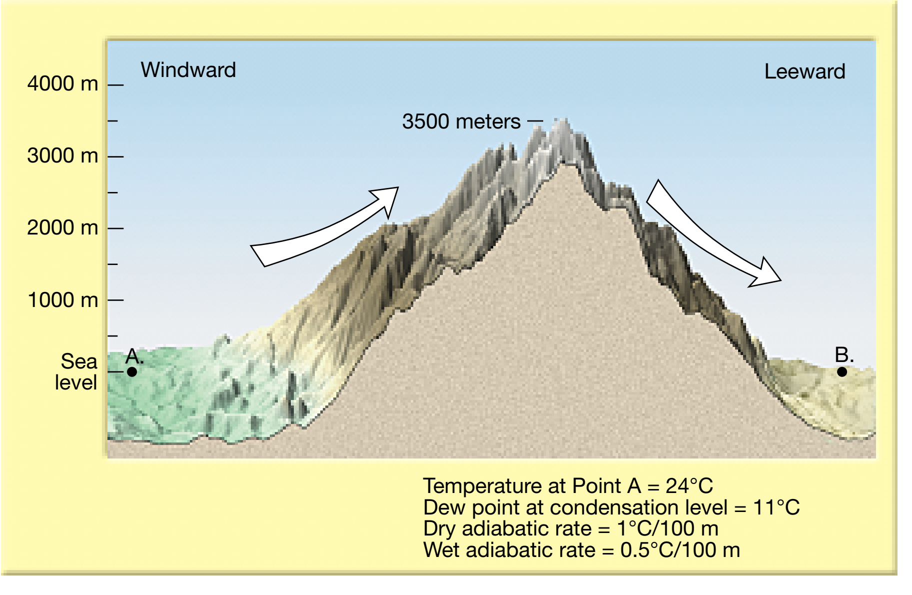


FIGURE 4-30 Orographic lifting problem.

Atmospheric Science Online

The Atmosphere 10e web site uses the resources and flexibility of the Internet to aid in your study of the topics in this chapter. Written and developed by meteorology instructors, this site will help improve your understanding of meteorology. Visit **http://www.prenhall.com/lutgens** and click on the cover of *The Atmosphere 10e* to find:

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