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CHAPTER 4

Water in the Atmosphere



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After completing this chapter, you should be able to:

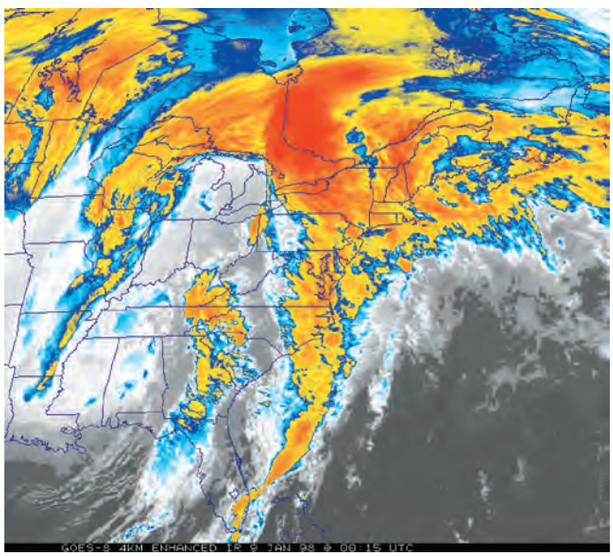
- Define *saturation* and its importance in the atmosphere
- Explain why there can be more water vapor in warm air than cold air and how this affects the atmosphere
- Describe how clouds and precipitation form
- Identify the major cloud and precipitation types and explain the significant differences among them

Introduction

In early January 1998, the Canadian province of Quebec was literally crushed by water falling from the sky. For a solid week, water vapor slid upward over the northeastern United States and central and eastern Canada. The vapor turned to clouds (yellow and orange areas in satellite picture), and precipitation dropped out of the clouds. Rain fell on Quebec, but the temperatures at the ground were cold enough for the rain to freeze on contact with roofs, roads, trees, and towers.

The freezing rain coated Canada with as much as 10 centimeters (4 inches) of ice. The ice clung to everything it touched. Its heavy weight broke 30,000 wooden power poles. Incredibly, the ice crumpled 1000 steel electrical transmission towers into pretzel shapes and destroyed them (see photo). About 1.5 million people in Canada lost electricity, and half of those still were without power a month later. One fifth of the nation's workers could not get to work easily in the wake of the storm. Not even the ageless Rolling Stones were immune. The ice crushed the roof at Olympic Stadium and caused them to cancel their Montreal concert.

In this chapter, we explore water in the atmosphere in all its phases: water vapor, clouds, and precipitation. We will explain how fog, clouds, and precipitation form. We will also learn how slightly different weather conditions can turn a cold rain into pellets of ice or into a destructive freezing rain that can paralyze half a nation.



CIMSS, University of Wisconsin-Madison



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Evaporation: The Source of Atmospheric Water

Water is, near the surface, the atmosphere's most abundant trace gas. How does water enter the atmosphere? **Evaporation** puts it there. As we learned in Chapter 1, evaporation is the process by which water is converted from liquid form into its gaseous state, water vapor. Evaporation occurs constantly over the surface of the Earth. When water molecules at the surface of liquid water gain enough energy to escape as vapor into the air above, evaporation results. As we learned in Chapter 2, it takes a lot of latent heat energy to change liquid water to vapor. Evaporation therefore occurs more rapidly over warmer surfaces, which supply water molecules with enough energy to escape into the atmosphere. Evaporation is also greater when the atmospheric pressure is low, the wind speed is high, and there is relatively little water vapor already in the air.

To better understand evaporation, let's consider the following example. Put some liquid water in a closed container. Keep the container at a constant temperature and pressure. Initially the container has only liquid water in it (Figure 4.1, *A*). Some individual molecules in the liquid water will have more (and some will have less) kinetic energy than the average. For instance, a water molecule in the liquid phase might gain kinetic energy considerably above the average because of several rapid collisions with neighboring molecules. Now imagine such a molecule at the liquid's surface, the boundary between the water and the air. If it has enough kinetic energy to overcome the attractive force of nearby molecules and is moving toward the air, it may escape from the liquid (Figure 4.1, *B*).

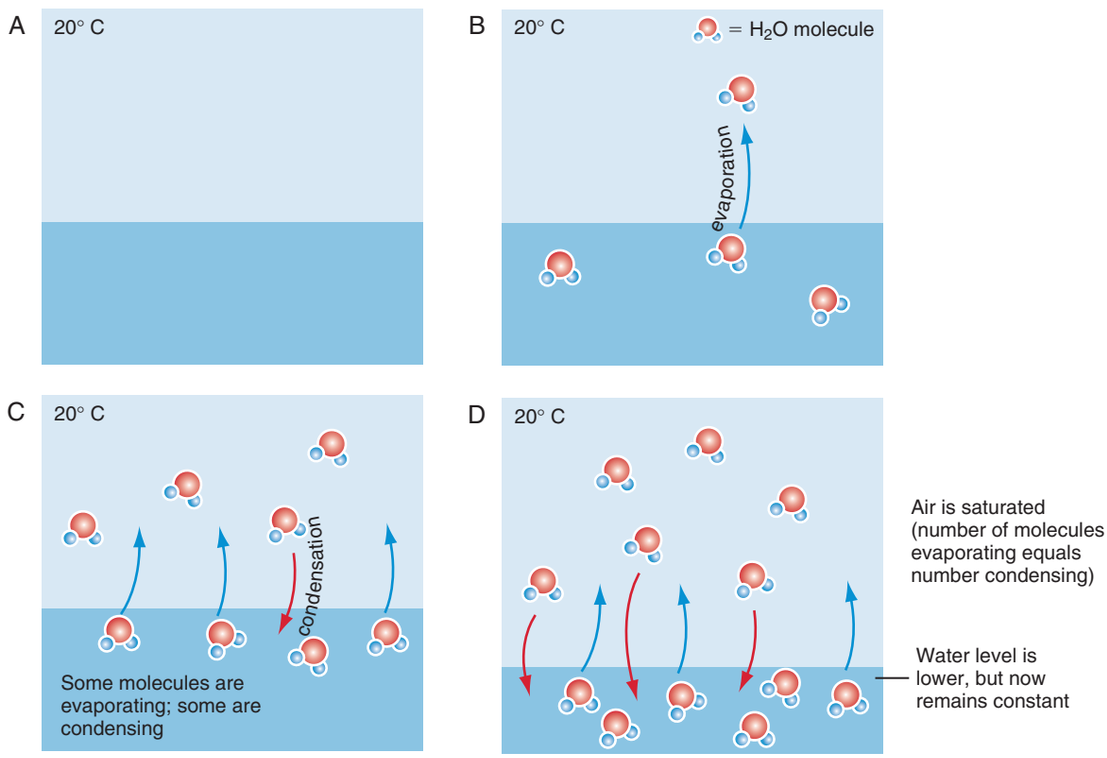


Figure 4.1

The sequence of events that leads to saturation of air. In **A,** *initially dry air lies above water at 20° C. In* **B,** *evaporation into the air begins, and in* **C,** *condensation back to the water occurs. However, evaporation exceeds condensation and the number of water molecules in the air increases. The air is saturated in* **D,** *where the number of evaporating and condensing water molecules is equal.*

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As time goes on, some water molecules at the liquid surface will be escaping and evaporating. Others will condense (Figure 4.1, C). Eventually, because the container is sealed, the number of molecules leaving the surface of the liquid will be the same as the number entering. This means there will be no net change in the number of molecules in the liquid or the vapor phase (Figure 4.1, *D*).

A situation in which there is no net change is described as being in *equilibrium*. When the number of molecules leaving the liquid is in equilibrium with the number condensing, the air above the surface is **saturated**—that is, the rate of return of water molecules is exactly equal to the rate of escape of molecules from the water. As we will see, the concept of saturation is central to understanding the formation of clouds and precipitation.

Counting the number of molecules in the container above the water is one way to measure the amount of water. There are several methods of measuring the amount of water vapor in the atmosphere that are easier to carry out and do not require counting molecules. We look at these different methods in the following.

Measuring Water Vapor in the Air

Why do we need to measure the amount of water vapor in the atmosphere? There are several reasons:

1. The change of phase of water is an important energy source for storms, atmospheric circulation patterns, and cloud and precipitation formation.
2. Water vapor is the source of all clouds and precipitation. The potential for cloud formation and dissipation depends on the amount of water vapor in the atmosphere.
3. The amount of water in the atmosphere determines the rate of evaporation. Rates of evaporation are important to weather and to many forms of plant and animal life, including humans
4. Water vapor is a principal absorber of longwave radiant energy. It is the most important greenhouse gas.

News reports of current weather conditions often include the dew point temperature and the relative humidity. These are just two of several ways to express the amount of water vapor in the atmosphere. Each way has advantages and disadvantages. In this section, we will discuss four different methods of representing the amount of water vapor in the atmosphere: mixing ratio, vapor pressure, relative humidity, and dew point/frost point.

Mixing Ratio

One way of expressing the amount of water vapor in the atmosphere is the ratio of the weight of water vapor to the weight of the other molecules in a given volume of air. This is the **mixing ratio.** The unit of mixing ratio is grams of water vapor per kilogram of dry air. Typical values of the mixing ratio near the surface of the Earth range between less than 1 gram per kilogram in polar regions to more than 15 grams per kilogram in the tropical regions. Because the surface of the Earth is the source of water vapor for the atmosphere, the mixing ratio generally decreases as you get farther above the surface (Figure 4.2).

Mixing ratio is an absolute measure of water vapor. This means that it is proportional to the actual number of water molecules in the air. Adding or removing water vapor molecules from a fixed volume of air changes its mixing ratio. Evaporating water into the volume increases the mixing ratio. Since the mixing ratio has to do only with the weight of water vapor relative to the total weight of an air mass and because the total mass and total number of molecules remain unchanged, cooling the air or expanding the air has no effect on the value of the mixing ratio.

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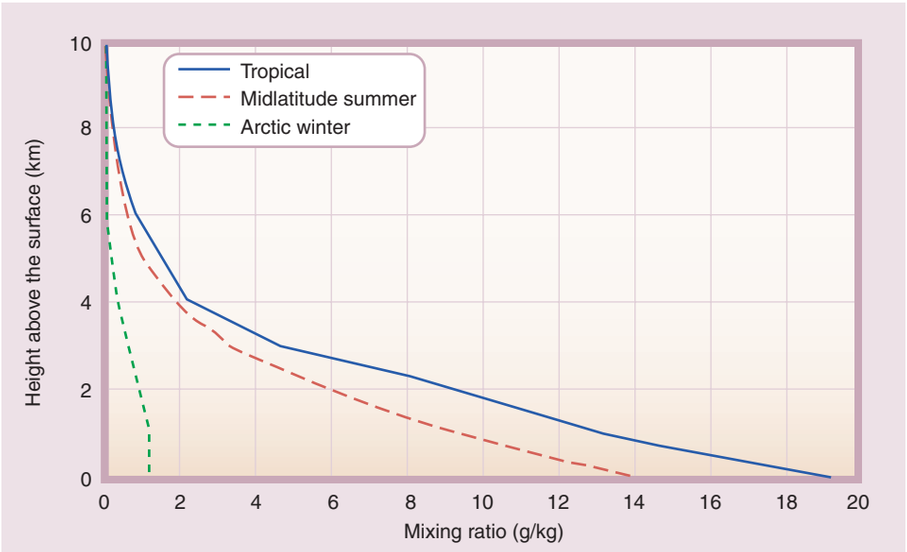


Figure 4.2
Atmospheric water vapor comes from the surface of the Earth. This explains why values of mixing ratio usually decrease with altitude even at different regions of the world and during different seasons.

Vapor Pressure

Gas molecules exert a pressure when they collide with objects. The atmosphere is a mixture of gas molecules, and each type of gas contributes its part of the total atmospheric pressure. The pressure the water molecules exert is another useful method of representing the amount of water vapor in the atmosphere. The pressure caused by these water vapor molecules is called the **vapor pressure.** Atmospheric vapor pressure is expressed in millibars (mb).

As we learned in Chapter 1, water vapor is at most only 4% of the total atmosphere. The average surface pressure as a result of all atmospheric gases is approximately 1000 mb. Therefore the vapor pressure attributable to water vapor alone is never more than about 4% of 1000 mb, or 40 mb.

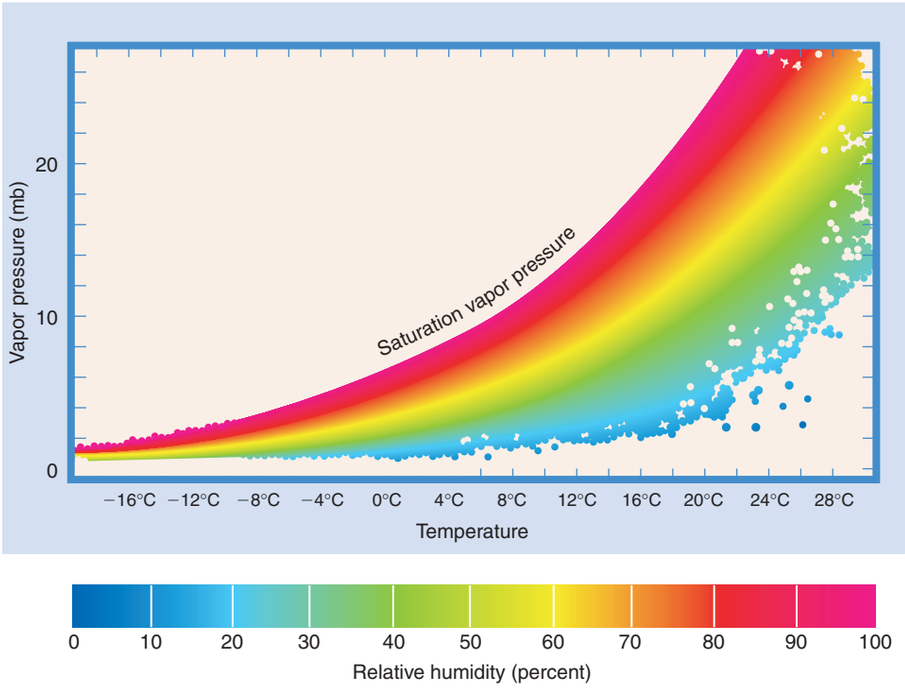
A variety of factors can change the vapor pressure. The higher the temperature, the greater the average kinetic energy of the molecules. Both the total pressure and the vapor pressure increase with temperature. Increasing the number of water vapor molecules in a specific volume of air will also raise the vapor pressure. When water evaporates into a volume of air, both the vapor pressure and the mixing ratio increase. However, if the air cools, the vapor pressure decreases along with the total air pressure, but the mixing ratio remains constant. Atmospheric scientists use vapor pressure to express the amount of water in the atmosphere when they discuss the formation of clouds.

When air is saturated (as in Figure 4.1, *D*), the pressure exerted by the water vapor molecules is called the **saturation vapor pressure.** Saturation vapor pressure in the atmosphere is reached whenever the atmospheric water vapor exerts a pressure equal to what the saturation vapor pressure would be at that particular temperature in a closed container.

Figure 4.3 reveals several facts about vapor pressure and saturation vapor pressure. The data in Figure 4.3 represent more than 6 years of hourly observations of vapor pressure and temperature on a ridgetop 80 kilometers (50 miles) north of New York City. What does this graph tell us? The lack of observations in the top left part of the graph implies that vapor pressure cannot exceed a certain value for a given temperature. The maximum value at each temperature is the saturation vapor pressure for that temperature. Following the maximum (*pink*) values from left to right, you see a curve that swoops upward rapidly as the temperature increases.

MeteorologyNow™ Click the *Meteorology Interactive* "Moisture Graph" to learn how moisture affects static stability

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Active Figure 4.3 MeteorologyNow™

Observations of vapor pressure as a function of temperature on a ridgetop at Black Rock Forest along the lower Hudson River in New York. The observations were made hourly from December 1994 through mid-April 2001. Relative humidity is indicated by the color-coding. Notice how the highest observed values of vapor pressure at each temperature form an arc that curves upward from left to right. This upper limit on vapor pressure at each temperature is the saturation vapor pressure.

(Based on data from the Open Lowland environmental monitoring station at Black Rock Forest, Cornwall, NY. Used by permission of the Black Rock Forest Consortium.)

This last point is the most important fact about saturation vapor pressure: *It increases rapidly as the temperature increases*. Why? As the temperature of water increases, the number of molecules with enough kinetic energy to evaporate from the water surface increases. Increasing the temperature also increases the number and speed of the water molecules in the vapor phase. As a result, more molecules move at greater speed and exert a higher pressure.

It is often said that "warm air holds more water vapor than cold air," but this is a misleading simplification. This saying implies that warm air expands and has more room for water vapor, which is incorrect. Instead, the amount of water vapor in the air is, as we have discussed here, an equilibrium between the air and the surface beneath it. Moreover, the saying is sometimes false—for example, when comparing a hot dry desert to a cold moist coastline. It is more accurate to say that a saturated parcel of warm air will contain many more water vapor molecules than a saturated parcel of cooler air.

Relative Humidity

Neither the vapor pressure nor the mixing ratio tells us how close the air is to being saturated. The ratio of the actual vapor pressure exerted by molecules of water vapor versus the saturation vapor pressure at the same temperature indicates just how close the air is to saturation. This ratio is called the *saturation ratio*. Multiplying the saturation ratio by 100% yields the **relative humidity.**

$$\text{Relative Humidity} = 100\% \times \frac{\text{Vapor Pressure}}{\text{Saturation Vapor Pressure}}$$

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Relative humidity describes how far the air is from saturation. Saturated air has a relative humidity of 100% because the vapor pressure equals the saturation vapor pressure. In Figure 4.3, the pink colors indicate relative humidities near 100%, and they are also the maximum values of vapor pressure for a particular temperature. This implies that when the relative humidity is close to 100%, the vapor pressure is very close to the saturation vapor pressure—which is exactly what the formula tells us. A relative humidity of 50% (*light green* in Figure 4.3) tells us the vapor pressure is half that required for saturation.

The amount of relative humidity also affects the rate of evaporation. At the same pressure and temperature, water evaporates more slowly in air that has a high relative humidity and more quickly in air that has a low relative humidity. This fact is of prime

BOX 4.1 Atmospheric Moisture and Your Health

Water in the atmosphere can be a killer in all of its phases: vapor (humidity), liquid (fog and rain), and ice (freezing rain and ice storms). The victim can be a National Football League star—or you.

On August 1, 2001, 27-year-old Minnesota Vikings football player Korey Stringer died just hours after practicing with his team. How could such a physically fit young athlete, who was selected by his peers to play in the Pro Bowl the previous year, die so suddenly?

When our bodies get hot we cool down by sweating. It is not the sweating that cools our bodies; it is the evaporation of the sweat. If the air has a high vapor pressure, then the rate of evaporation is reduced. This hampers the body's ability to maintain a nearly constant internal body temperature. This is why we are uncomfortable on hot, muggy days. Like an engine without proper ventilation, we are overheating.

When athletes practice for hours in summertime heat and humidity, their bodies can overheat to the point that organs can fail and death can occur. This is called heatstroke. For example, Stringer's temperature when he reached the hospital was above 108.8° F, more than 10° above normal.

The *apparent temperature index* or *heat index* (see the accompanying table) indicates how hot it *feels*. It is expressed as a function of air temperature and the relative humidity. R.G. Steadman developed this index in 1979. When the temperature is high but the relative humidity is low, the heat index is less than the actual temperature. This is because cooling by evaporation of sweat is very efficient in these situations. However, high relative humidities prevent evaporation and make it seem hotter than it really is. In these cases, the heat index is greater than the actual temperature.

A combination of high temperature and high humidity leads to extreme heat indices as much as 40° F above the actual air temperature. In these situations, exercising outside can be fatal. For example, the Vikings practiced in Mankato, Minnesota, in the morning to avoid daytime maximum temperatures. The air temperature at noon in Mankato on the day of Stringer's last practice was 89° F, which does not sound particularly unusual for summertime. However, the dew point was a sultry 80° F, giving a relative humidity of 75%. The high humidity in Mankato, combined with the temperature, yielded a heat index of 106° F. As the table shows, heatstroke was indeed a possibility at the Vikings' practice that morning.

Athletes are not the only ones at risk from high heat index values. Prolonged periods of very high temperatures in association with high humidities can be extremely dangerous even to those who are not exercising at the time, as we will discuss in Chapter 9 .

Fog can be a killer, too. Each year 680 fatalities occur in the United States as a result of traffic accidents during which fog is present. A combination of high speed and low visibility is often to blame. On December 27, 1996, fog caused a 54-vehicle series of pileups on the Sunshine Skyway Bridge over Tampa Bay, Florida. The wrecks killed one person and snarled traffic for 7 hours. Fog also played a key role in the United States' worst modern passenger train wreck. A towboat lost in fog bumped a railroad bridge near Mobile, Alabama, early on September 22, 1993. The bridge was pushed out of alignment, causing an Amtrak train to derail as it crossed the bridge a few minutes later. The train plunged into the water, killing 47 people.

Rain also kills. Up to 20% of all fatal highway accidents occur on wet pavement. A little rain after a dry spell, combined with the built-up residue of oil on roads, turns concrete and asphalt into a slick and oily mess. Cars can "hydroplane" on the wet surface and skid out of control because of reduced friction between the tires and road surface.

Ice storms consisting of freezing rain and sleet are perhaps the most dangerous of all. Driving in an ice storm is life-threatening, much more uncontrollable than in rain. Even off the roadways, you can easily fall and break bones on slippery steps or a sidewalk.

In 1999 automobile accidents attributed to bad weather caused approximately 5000 deaths and 500,000 injuries in the United States, according to the National Highway Traffic Safety Administration. Whether exercising in hot humid weather or driving in fog, you should give careful consideration to possible threats to your safety related to the water in your environment.

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importance to the public because high humidity makes perspiration an inefficient way of removing heat by evaporation. This can lead to uncomfortable and even life-threatening conditions (Box 4.1). Relative humidity is more generally an important indicator of the rate of moisture and heat loss by plants and animals.

Relative humidity can change in response to a wide range of circumstances. For example, in the case of a constant volume of air at a constant temperature, changing the vapor pressure changes the relative humidity. Why does the relative humidity change in this case? Adding water molecules to a fixed volume of air increases the vapor pressure but has no effect on the saturation vapor pressure because the temperature has not changed. The saturation ratio is changed, and so is the relative humidity.

BOX 4.1 *(continued)*

Relative humidity (%)	Air temperature (°F)										
	70	**75**	**80**	**85**	**90**	**95**	**100**	**105**	**110**	**115**	**120**
0	64	69	73	78	83	87	91	95	99	103	107
10	65	70	75	80	85	90	95	100	105	111	116
20	66	72	77	82	87	93	99	105	112	120	130
30	67	73	78	84	90	96	104	113	123	135	148
40	68	74	79	86	93	101	110	123	137	151	
50	69	75	81	88	96	107	120	135	150		
60	70	76	82	90	100	114	132	149			
70	70	77	85	93	106	124	144				
80	71	78	86	97	113	136					
90	71	79	88	102	122						
100	72	80	91	108							

- Great risk to health, heatstroke imminent.
- Risk of heatstroke.
- Prolonged exposure and physical activity could lead to heatstroke.
- Prolonged exposure and physical activity may lead to fatigue.

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Changing the saturation vapor pressure also changes the relative humidity. As shown in Figure 4.3, the saturation vapor pressure decreases rapidly when the temperature of the air decreases. Therefore, a decrease in temperature results in an increase in the relative humidity, and increasing the temperature decreases the relative humidity.

The effect of temperature on relative humidity is illustrated in Figure 4.4, which is based on observations at a national monument in New Mexico. For every season of the year, relative humidity peaks around sunrise and is at its lowest in mid-afternoon. Why? The saturation vapor pressure is lowest at sunrise, when the temperature is lowest. Therefore, without a change in the actual moisture content of the air, the relative humidity will be highest at sunrise. Conversely, daytime heating raises the temperature and the saturation vapor pressure, reducing the relative humidity. This seesaw pattern of temperature and relative humidity is seen during mostly clear, relatively calm, and precipitation-free conditions, which exist most of the time in New Mexico.

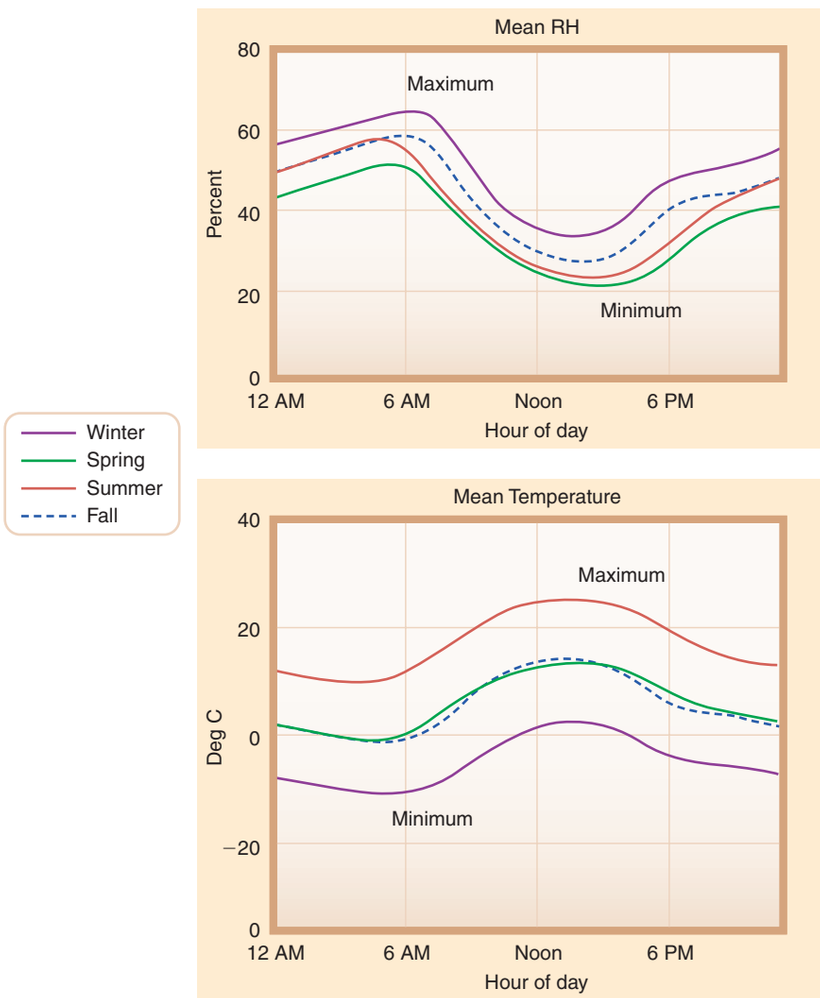


Figure 4.4

A climatology of hourly temperature and relative humidity, as observed at Bandelier National Monument, New Mexico, from October 1988 through May 1999. In the usually dry, clear climate of New Mexico, the daily cycle of relative humidity is exactly the opposite of the temperature cycle, with a peak at sunrise and a minimum at mid-afternoon. The same daily cycle of relative humidity is observed at any location during a mostly calm, clear, dry day.

(Adapted from http://vista.cira.colostate.edu/improve/Data/GraphicViewer/Diurnal.htm).

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The effect of temperature on relative humidity explains why regions with cold winters may have very low indoor relative humidities. As cold outside air finds its way into a building it is eventually heated, and this greatly decreases the air's relative humidity. Explore this for yourself using our Exploring Humidity learning applet.

MeteorologyNow™ Click the *Atmospheric Exploration* "Exploring Humidity" to explore why, for example, homes in cold climates are so dry in winter.

To summarize: *Adding water vapor, cooling the air, or both increases relative humidity; removing water vapor, warming the air, or both decreases relative humidity.*

Dew Point/Frost Point

When air is saturated, the vapor pressure is equal to the saturation vapor pressure. The air cannot contain any more moisture. If the vapor pressure is greater than the saturation vapor pressure, the relative humidity exceeds 100%. Ordinarily, this is not possible in the atmosphere. The excess moisture must condense out of the air until the relative humidity is once again reduced to 100%. This condensed water is called **dew** (Figure 4.5).

Dew forms when moisture is added to the air, when the air is cooled, or when a combination of both moistening and cooling occurs. The most commonplace occurrences of dew are caused by cooling. An everyday example of dew is the moisture that forms on the outside of an ice-cold drink glass. Overnight cooling of the air near the ground causes morning dew on grass, car windshields, and spider webs.

The temperature to which air must be cooled to become saturated without changing the pressure is called the **dew point.** The dew point temperature is determined by keeping the pressure fixed because changing the pressure affects the vapor pressure and therefore the temperature at which saturation occurs.

To know how close the air is to saturation, we need to know the dew point and the air temperature. *The closer the dew point is to the air temperature, the closer the air is to saturation.* When the dew point equals the air temperature, the air is saturated, so the dew point temperature cannot be greater than the air temperature. The temperature difference between the air and the dew point is called the **dew point depression.** Table 4.1 compares all the different measures of water vapor in the atmosphere for two different temperatures and two different relative humidities.

If temperatures are below freezing, saturation or moistening of cooled air will lead to deposition (Chapter 2) rather than condensation. The ice crystals that form are called



Steve Ackerman

Figure 4.5
When the temperature of the air around this web cooled to the dew point temperature, dew formed, making the web more visible.

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TABLE 4.1 **Various humidity quantities for two air temperatures and two relative humidities for an atmospheric pressure of 1000 mb**

Temperature	−10° C (14° F)		20° C (68° F)	
Relative humidity	25%	75%	25%	75%
Mixing ratio (g/kg)	0.45	1.35	3.67	11.15
Vapor pressure (mb)	0.72	2.16	5.87	17.60
Saturation vapor pressure (mb)	2.88	2.88	23.47	23.47
Dew point temperature	−26.2° C (−15.2° F)	−13.5° C (7.8° F)	−0.5° C (31.3° F)	15.6° C (60.1° F)
Dew point depression	16.2° C	3.5° C	20.5° C	4.4° C

frost. The temperature to which air must be cooled at a constant pressure to cause frost to form (normally below 0° C [32° F]) is called the **frost point.** Dew may form and then freeze if the temperature falls below freezing, forming *frozen dew*. Frozen dew is different than frost. Frozen dew first condenses as liquid water before freezing, instead of becoming ice via deposition, as in the case of frost.

We can use the energy budget concepts from Chapters 2 and 3 to explain dew and frost. Dew and frost form on objects in air close to the ground, such as blades of grass. Whether a blade of grass cools below the dew or frost point is a function of its energy gains and losses. At night, a blade of grass loses energy by emission of longwave radiation while gaining energy by absorbing the longwave radiation emitted from surrounding objects. Under clear nighttime skies, objects near the ground emit more radiation than they receive from the sky, and so a blade of grass cools. If the temperature of a grass blade falls below the dew or frost point, dew or frost will form on the grass.

This explains why frost forms in an open field but not under a tree (Figure 4.6). Trees emit more radiation toward the ground than does the clear sky. Energy gains of the grass in the open field are less than those of the grass under the tree. The grass in the open field cools faster and reaches the frost point before the grass blades under the tree.

The dew point is useful in forecasting minimum temperatures. On a clear, calm night, the temperature will often drop to near the dew point. This is because condensation releases energy, and this energy release counteracts cooling below the dew point.



Steve Ackerman

Figure 4.6
Frost, ice crystals formed by deposition of water vapor on subfreezing surfaces, will form in open fields before forming under a tree. In the background, steam fog is forming over the water.

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Formation of frost and dew are examples of phase transitions between the gas phase of water and its solid and liquid states. These transitions are vital to understanding how clouds form, which we now examine in detail.

Condensation and Deposition: Cloud Formation

Clouds are the atmosphere's equivalent of movie stars, instantly recognizable worldwide and the subject of endless curiosity. Children and poets look for shapes in the clouds. Parents struggle to explain how something so large can keep from falling out of the sky but then can disappear in a matter of minutes.

The key to understanding clouds is water. Clouds, from the fair-weather wisps to the mightiest thunderstorms, are composed of nothing more than tiny 20-micron-sized particles of liquid water called **cloud droplets** and particles of ice called **ice crystals.** However, the formation and growth of these particles is one of the most complicated aspects of weather and climate. Before we can look at clouds, we must first examine in some detail how they are made.

Solute and Curvature Effects

As a volume of unsaturated air cools, its relative humidity increases. If the air is sufficiently cooled, the temperature equals the dew point and the relative humidity equals 100%. Based on what we have learned so far, condensation should occur at this point, forming a cloud. But cloud droplets can actually form at relative humidities other than 100%. Why?

Over the oceans and seas, waves and winds add salt to the mix of atmospheric components. When salt dissolves in water, the salt particles become dispersed among the water molecules. Salt particles dissolved in water attract water molecules even more strongly than neighboring water molecules. The greater the concentration of salt, the more the rate of evaporation is reduced, all other things being equal.

The ability of dissolved salt to hold onto water molecules is called the **solute effect.** By suppressing evaporation, the solute effect enhances the growth of droplets by condensation, thereby allowing cloud formation at relative humidities much less than 100%. As the droplet grows, however, the solution becomes more dilute and the solute effect decreases.

In our previous discussion of evaporation, we discussed a single molecule near the edge of a flat surface of still water—but cloud droplets aren't flat surfaces. A molecule on any surface feels attracted to its neighbors, which attempt to keep it part of the water. A molecule on a curved surface such as a cloud droplet has fewer neighbors to attract it (Figure 4.7) and can therefore escape the fluid more easily.

As a result, even if air is saturated with respect to a flat surface of water, it may be unsaturated with respect to a curved surface. This is called the **curvature effect.** This effect opposes the formation of small droplets by condensation. As a result, the relative humidity must be higher than 100%—a condition known as **supersaturation**—for cloud formation to occur.

It is surprisingly difficult to form a water droplet out of air that contains only water vapor. It takes a relative humidity of more than 200% for water vapor molecules to form a tiny cloud droplet of pure water. This is because a tiny droplet has a strongly curved surface, but relative humidities this high are never observed in the atmosphere. So how do liquid droplets form from pure vapor? This question leads us to the subject of nucleation.

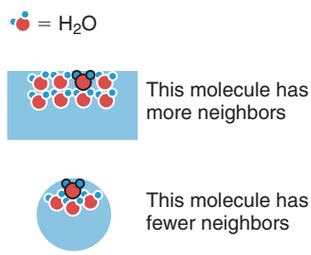


Figure 4.7
The smaller the drop, the more curved the surface, reducing the number of neighbors for each water molecule at the surface. This curvature effect makes it easier for small drops to evaporate.

Nucleation

Droplets form around particles. The initial formation of a cloud droplet around any type of particle is called **nucleation.** There are two types of nucleation: homogeneous and heterogeneous nucleation. In **homogeneous nucleation,** the droplet is formed only by water

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molecules. Homogeneous nucleation requires that enough water molecules bond together to form a cluster, or particle, that then acts as a nucleus for further condensation. Water-only bonding only works if the water molecules have low kinetic energy. If the kinetic energy of the molecules is too high, the cluster cannot form. For this reason, homogeneous nucleation only occurs at temperatures colder than −40° C (−40° F).

You can see homogeneous nucleation for yourself when you open a chilled bottled beverage that has very clean air in the bottle's neck. By removing the cap, you allow the air inside the neck to expand adiabatically and cool rapidly, but temporarily, to −40° C. The smoky cloud in the neck of the bottle is the result of homogeneous nucleation.

We learned in Chapter 1 that temperatures are as low as −40° C only in the upper troposphere, close to the stratosphere. Clouds usually form in much warmer air. Therefore, most clouds must develop through a different process. **Heterogeneous nucleation** occurs when small, nonwater particles serve as sites for cloud droplet formation. The particles are usually aerosols such as those we studied in Chapter 1. The aerosols that assist in forming liquid droplets are called **condensation nuclei.**

In the next sections, we will first consider the formation of liquid droplets around condensation nuclei and then address how ice crystals form around ice nuclei.

Condensation Nuclei

There are two types of condensation nuclei: hygroscopic and hydrophobic. **Hygroscopic nuclei** dissolve in water, and **hydrophobic nuclei** do not. Nucleation is more favorable on hygroscopic ("water-seeking") nuclei. Droplet formation can occur on hygroscopic nuclei even when the relative humidity is below 100% because the solute effect reduces the rate of evaporation. Hydrophobic ("water-repelling") nuclei resist condensation but can form droplets when relative humidities are near 100%.

There are plenty of condensation nuclei in the atmosphere in the form of dust, salt, pollen, and other small particles. The surface of the Earth is the major source of aerosols. The concentration of condensation nuclei is therefore usually greatest near the surface and decreases with altitude. In general, there is no lack of condensation nuclei for forming water droplets. Polluted cities have more condensation nuclei than wilderness environments. Over the oceans the air has fewer condensation nuclei than over land. Many of the nuclei over the oceans also contain salt thrown from waves, making them hygroscopic nuclei.

Ice Nuclei

When ice crystals form, water molecules cannot deposit onto the crystal haphazardly, as they can when condensing onto an existing water droplet. The molecule must fit into the shape of the crystal. **Ice nuclei,** the particles around which the ice crystals form, are important in the beginning stages of ice crystal formation. The ice nuclei make it easier for deposition to occur. Ice particles can form in four ways: deposition nucleation, freezing nucleation, immersion nucleation, and contact nucleation.

In **deposition nucleation,** ice forms from vapor by deposition onto the ice nucleus when the air is supersaturated with respect to ice. This happens most often on particles, such as clay, that have a molecular geometry resembling the molecular structure of ice. This geometry helps water molecules in the surrounding air to align in the proper molecular structure for forming ice when they deposit on the surface of the nuclei.

Liquid water at a temperature below 0° C is referred to as **supercooled water. Freezing nucleation** is the process by which a supercooled drop freezes without the aid of a nonwater particle.

The existence of supercooled water requires some explanation. There is a big difference between freezing a small water droplet and freezing a larger body of water. The freezing point of a large body of water (such as the water in an ice tray) is 0° C at standard pressure. However, a 1-millimeter diameter droplet will generally not freeze until the temperature falls below −11° C (12.2° F). A tiny droplet, but not a larger body of water, can be supercooled.

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How can this happen? For ice to form all the water molecules must align in the proper crystal structure. First a few molecules align, and then the rest quickly follow, turning the liquid into a block of ice. The larger a volume of water is as its temperature falls below freezing, the higher the chances that a few of the molecules will line up in the proper manner. In a small volume of water the chance that some of the molecules will align in the correct structure is reduced, simply because there are fewer molecules. For this reason, 0° C is more accurately called the *melting* point, not the freezing point, of water.

In **immersion nucleation** the nucleus is submerged in a liquid drop. Once the drop reaches a given temperature, the immersed nucleus allows the supercooled liquid to rapidly align in the crystalline structure of ice, causing the drop to freeze.

Ice nuclei may also collide with supercooled drops. The drop freezes immediately on contact with the ice. This is referred to as **contact nucleation.**

Cloud Particle Growth by Condensation and Deposition

Once a cloud particle forms, it can grow if the air around it is saturated. If the particle is a liquid droplet, and if the vapor pressure of the air is greater than the vapor pressure just above the surface of the particle, it will continue to grow by condensation. If the particle is ice, it grows by deposition.

Growth by condensation and deposition produces droplets, but they are small. It takes a long time to create droplets large enough to fall as precipitation. We look at precipitation particle growth a little later in this chapter. First, we examine how growth by condensation produces fog, a cloud at the ground.

Fog Formation

The air in contact with the ground can become saturated if it cools or when water from the surface evaporates into it. Water vapor then condenses on cloud condensation nuclei to form a suspension of tiny water drops. This is a cloud at the ground, which we call **fog.**

The formation of heavy fog often reduces visibility to the point where certain modes of transportation become hazardous (Figure 4.8). The appearance of fog on highways can trigger chain-reaction accidents involving scores of vehicles. Fog also contributed to the famous collision between the *Titanic* and an iceberg. In early December 1952, a fog in London became so thick (partly because of pollution) that people walked into canals and rivers because they could not see the ground.

The distribution of heavy fog over the continental United States is shown in Figure 4.9. Heavy fog in Alaska, Hawaii, and Puerto Rico occurs on fewer than 10 days per year. Fog is most common in the Appalachian Mountains and near bodies of water, especially along the northwest and northeast coasts.

Fogs are named for the ways in which they form. We explore four different types of fog below: radiation fog, advection fog, evaporation fog, and upslope fog.



Steve Ackerman

Figure 4.8
Fog consists of tiny water droplets that can reduce visibility to less than 1 kilometer (0.6 miles). Heavy fog can be a travel hazard.

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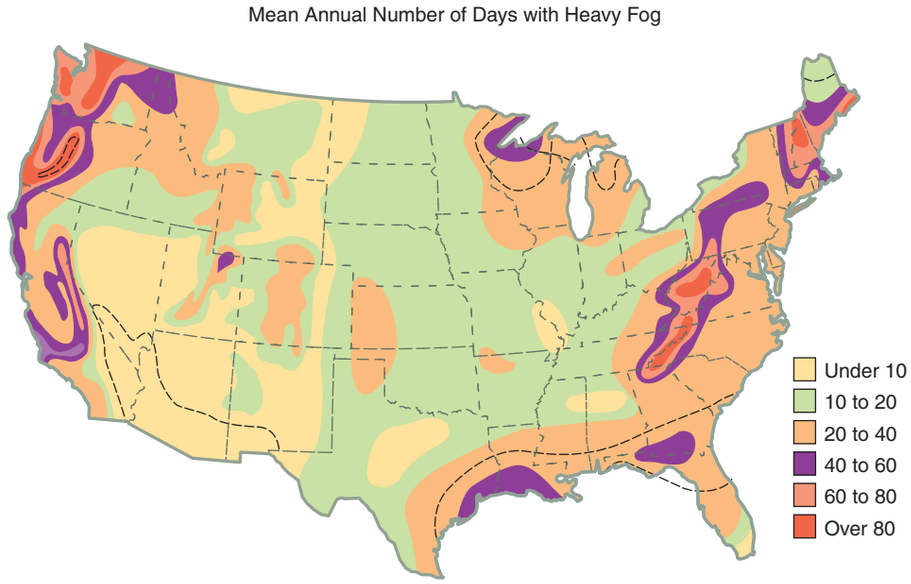


Figure 4.9
The mean annual number of dense fog days with visibility less than 300 meters across the United States.

Radiation Fog

Radiation fogs form in the same way that dew does. On clear, long nights the ground rapidly cools by radiation and the air just above the ground cools by conduction and radiation. As the temperature of the air drops, the relative humidity increases. Radiation fogs tend to develop on clear nights, when radiative cooling near the ground is more rapid. Light winds are also required because they can gently mix moist air near the ground. Winds that are too strong mix the air near the ground with the drier, warmer air above. This keeps the air near the surface from saturating.

Radiation fogs are common in autumn in river valleys and small depressions. The cold air sinks to the bottom of the valley, providing the cool air. Rivers and streams provide the water vapor needed to increase the relative humidity via evaporation. These fogs are often called *valley fogs* (Figure 4.10).

There are some rules for forecasting a radiation fog. If the dew point temperature is approximately 8° C (14° F) below the air temperature at sunset, and if the winds are predicted to be less than 9 kilometers per hour (5 knots), there is a good chance that a radiation fog will form during the night.

Advection Fog

When warm air is advected (blown horizontally) over a cold surface, the air near the ground cools because of energy exchanges with the surface. The relative humidity increases, and an **advection fog** may form (Figure 4.11).

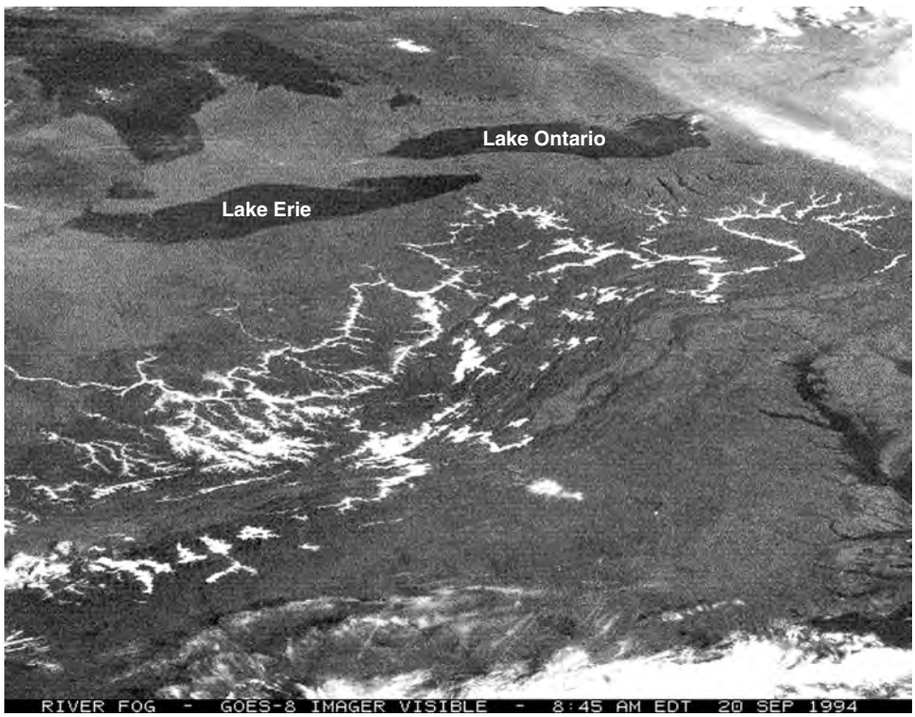
Advection fog is common off the coast of California as warm moist air over the Pacific is advected over the cold coastal waters. Off the East Coast, warm air over the warm Gulf Stream current may be advected over the colder coastal waters. Another foggy region is off the coast of Japan, where the cold water of the Oyoshio current meets the warm Kuroshio current. These fogs form at all times of the year and can last for more than a week.

Advection fogs can also occur when warm air flows from over the water to cooler land. Fog is common along the coast of the Gulf of Mexico during fall and winter. During these times saturation of the air occurs when warm moist air flows from the Gulf of Mexico over the cooler land. These types of fog are also common in New England and give London its reputation for fog.

Evaporation Fog

If you take a long hot shower, you may "fog up" the bathroom. Some of the warm water from the shower evaporates into the cooler bathroom air, moistening it to saturation and

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SSEC and CIMSS, University of Wisconsin-Madison

Figure 4.10
Satellite photograph on the morning of September 20, 1994, showing radiation fog (narrow white areas) in the Ohio River valley and its tributaries.

forming a fog. **Evaporation fogs** also occur in the vicinity of warm fronts and are sometimes called **frontal fogs**. These fogs form when water evaporates from rain that falls from warmer air above the ground into cold air near the surface. Frontal fogs only form after it has been raining for hours because it takes time for the evaporating drops to saturate the air. Similarly, it is difficult to fog up the bathroom by taking a short shower.

Evaporation fogs also form over lakes when much colder air moves over warmer water. The vapor pressure of the cold air is less than that of the air over the water. As a result, evaporation is rapid. This rapid evaporation saturates the air above the surface. The condensation further warms the air. This warmed air rises and mixes with the cold air above it, reaching saturation and causing more fog to form.

Evaporation fog over a lake gives the appearance of steam rising out of the water and is sometimes referred to as a **steam fog** (Figure 4.12). It is common over lakes during late autumn or early winter in the more northern mid-latitude regions of the globe. Steam fog is common when very cold air rushes over unfrozen waters.



NOAA

Figure 4.11
Advection fog is common off the coast of California near San Francisco as warm moist air over the Pacific is advected over the cold coastal waters.

Upslope Fog

Consider air rising over a mountain barrier. As the air rises, it expands and cools and the relative humidity rises. If the air becomes saturated, an **upslope fog** forms. Upslope fog is common in moist mountainous regions such as the Appalachian Highlands. This type of fog forms best when the air near the ground, before flowing upslope, is cool and moist. Therefore, it does not require much lifting before saturation occurs.

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Ralph Burnett

Figure 4.12
An aerial view of steam fog rising and moving downwind from a small lake in Australia. Using your knowledge of relative humidity plus the shadows in this photograph, can you determine what time of day this picture was taken?

Lifting Mechanisms That Form Clouds

Upslope fog occurs when air moves along a rising ground surface. In general, most clouds form when air cools to the dew point as a parcel of air rises vertically.

MeteorologyNow™ Click the *Meteorology Interactive* " Cloud Base Altitude" to explore the relationship between temperature, dew point, and the base of a cumulus cloud.

Figure 4.13 depicts four mechanisms that cause air to ascend. Air is lifted as it moves against a mountain range (Figure 4.13, *A*). The air cannot go through the mountain, and so it flows over the mountain. This is **orographic lifting.**

Other mechanisms can also cause clouds to form. For example, at the same pressure cold air is more dense than warm air. Fronts represent the boundaries between these air masses of different densities. As fronts move, **frontal lifting** occurs when less dense warm air is forced to rise over the cooler, denser air (Figure 4.13, *B*). Frontal lifting is common in winter. We study fronts in detail in Chapters 9 and 10.

During summer, **convection** is an important lifting mechanism. In summertime convection, solar energy passes through the atmosphere and heats the surface. The air near the surface warms, becomes less dense than the air around it, and rises (Figure 4.13, *C*).

The final mechanism, **convergence,** occurs when air near the surface flows together from different directions. When the air near the ground converges, or is squeezed together, it causes upward motion (Figure 4.13, *D*).

In each of these examples of lifting, the rising air creates an **updraft.** The updraft keeps the cloud particles suspended in mid-air despite the force of gravity that acts to bring them to the ground.

Certain atmospheric conditions are less favorable for cloud development than others. In the following, we examine the role of moisture in cloud development.

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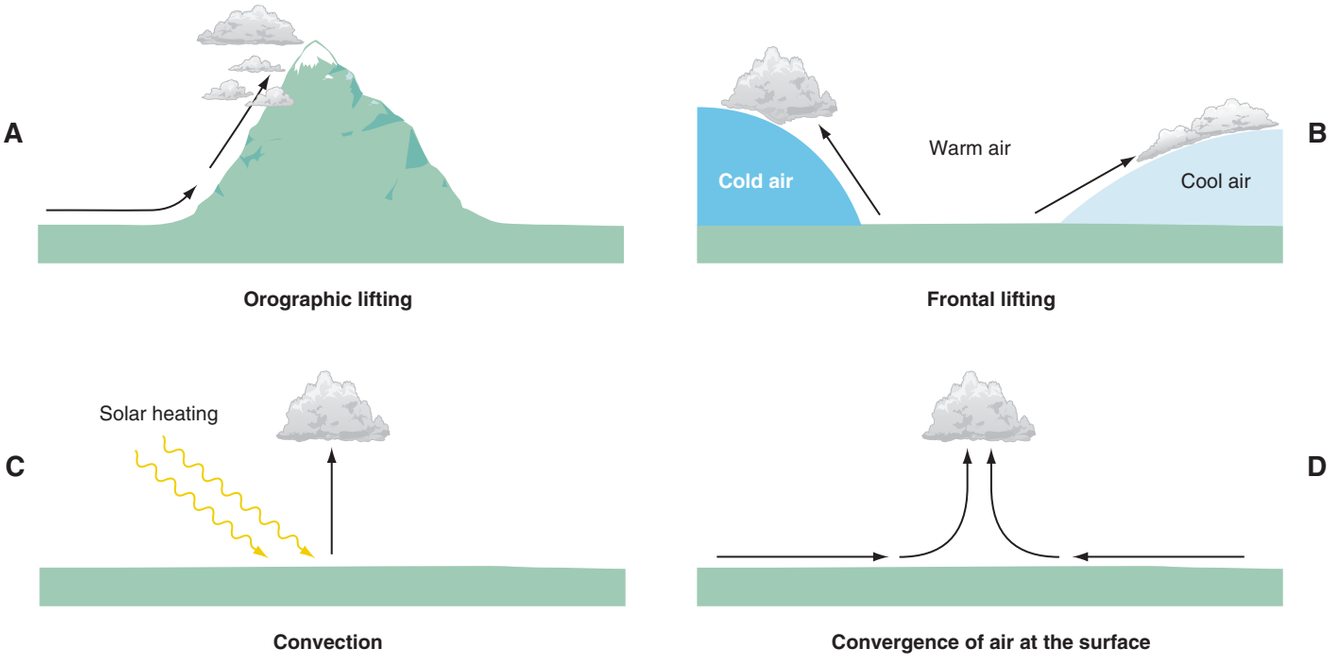


Figure 4.13
Four mechanisms that cause air to ascend and form a cloud.

Static Stability and Cloud Development

In Chapter 3 we introduced the concept of static stability. The basic question regarding stability is as follows: Will a rising air parcel keep on rising? To answer this question, we learned to compare the lapse rate of the environment to the dry adiabatic lapse rate.

A quick glance out the window during a thunderstorm, however, tells you that the atmosphere is not dry (i.e. unsaturated). Therefore, we have to modify our understanding of static stability to take water into account. This leads to two key new definitions: (1) a definition for a parcel's lapse rate that incorporates latent heating as a result of the phase changes of water, and (2) a definition for stability that includes this latent heating.

The Saturated Adiabatic Lapse Rate

A saturated parcel of air is one in which the air contains the maximum amount of water vapor possible; its relative humidity is therefore 100%. In saturated air, water molecules are changing phase from vapor to liquid or ice. As shown in Chapter 2, a phase change of water vapor to liquid water or ice releases energy, warming the parcel through latent heating.

This means that for an ascending moist parcel of air, two processes are going on at once. Expansion is cooling the parcel while condensation (or deposition) is warming the parcel by latent heating. The cooling process from expansion is always larger than the latent heating, so the parcel temperature decreases. The rate that the rising saturated air parcel cools is called the **saturated adiabatic lapse rate.**

Because heat is being added by the phase change of water vapor, the cooling rate of a rising saturated parcel is always less than the dry adiabatic lapse rate. The exact saturated adiabatic lapse rate for a given parcel depends on whether liquid or ice particles form and how much water vapor changes phase and, as a result, differs from one place and time to another. To simplify our discussion, in this text we will use a saturated adiabatic lapse rate of 6° C per kilometer.

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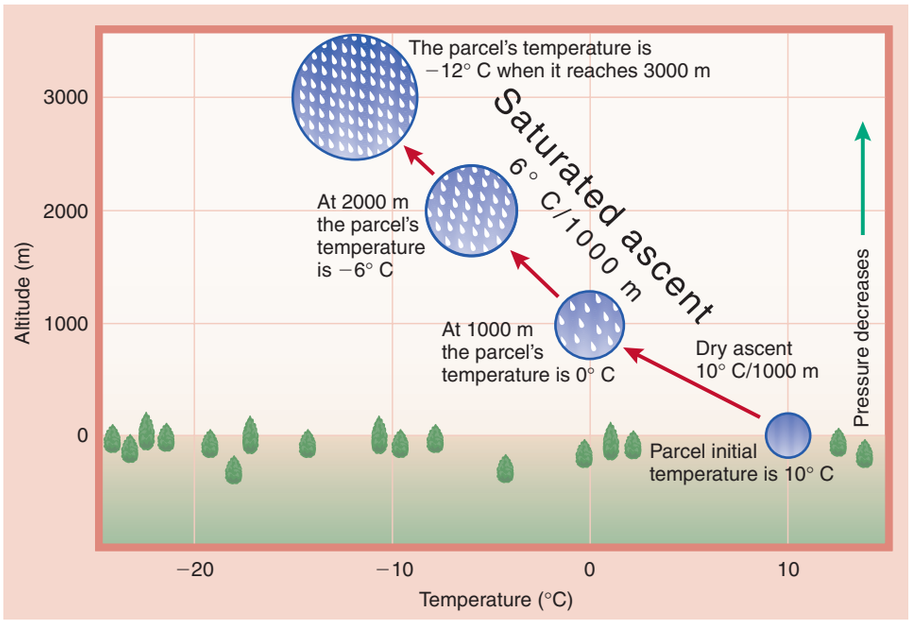


Figure 4.14

The ascent of a parcel of air that becomes saturated. The rate of cooling during ascent is less than the dry adiabatic lapse rate because of the warming as a result of latent heating, as water vapor molecules condense. A typical saturated adiabatic lapse rate is 6° C per kilometer, but the exact value varies. Compare this figure to Figure 3.17, which depicts the ascent of an unsaturated air parcel. How do the two differ?

To understand the saturated adiabatic lapse rate better, let's consider a rising parcel at the ground with an initial temperature of 10° C (50° F) (Figure 4.14). This is the same example we used in Chapter 3 to illustrate static stability with a completely dry parcel of air. This time, however, we will assume that our parcel becomes saturated at an altitude of 1000 meters.

As the parcel rises up to 1000 meters, it cools at the dry adiabatic lapse rate. When it reaches 1000 meters it has a temperature of 0° C (32° F). At this point the figure indicates that the water vapor in the parcel is condensing; in other words, the temperature and the dew point are equal at that altitude, and the relative humidity is 100%. This altitude is called the **lifting condensation level** (LCL) because it is the height at which water vapor in a rising parcel of air starts condensing. The bottoms of puffy clouds on sunny days are at the altitude of the LCL. The LCL varies with temperature and dew point, as you can demonstrate for yourself using the Cloud Base Altitude learning applet on the text's Web site.

If the parcel is warmer than its environment, or if it is being forced orographically up a mountain or front, it will continue rising. However, because the parcel is now saturated, as it rises above 1000 meters it will cool at the *saturated* adiabatic lapse rate. This is crucial! Up to now, our parcel of air has behaved just like the example in Chapter 3. But from here onward (and upward), there are big differences.

The parcel then rises to 2000 meters, where it has a temperature of −6° C (21° F), *not* −10° C as in the case of an unsaturated air parcel. The saturated air parcel is warmer than would be the case if it were unsaturated. Where does the added warmth come from? It comes from latent heating that is released as the water vapor condenses.

If the parcel continues at the saturated adiabatic lapse rate, what will its temperature be at an altitude of 3000 meters? Because the parcel is still saturated, it will cool to 6° less than −6° C, or −12° C (10° F). This is 8° C warmer than an unsaturated parcel that started at the ground with the same temperature.

The essential point is that ascending parcels that are saturated cool less quickly than do unsaturated parcels. This is the exact same thing as saying that the saturated adiabat-

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ic lapse rate is less than the dry adiabatic lapse rate. Next, how does this add to our understanding of static stability?

Conditionally Unstable Environments

In Chapter 3, we determined that an environment was either absolutely stable or absolutely unstable by comparing the environmental lapse rate to the dry adiabatic lapse rate. Now there is a third lapse rate to consider, however: the saturated adiabatic lapse rate. This creates a third possibility: that air parcels might be stable if they are "dry" (i.e., unsaturated) but unstable if they are saturated.

This third possibility depends on the *condition* of the air parcel—is it saturated or not? Therefore, the case where saturated air parcels are unstable, but unsaturated air parcels are stable, is called a **conditionally unstable** environment.

A conditionally unstable environment exists when its lapse rate is in between the saturated adiabatic lapse rate of about 6° C per kilometer and the dry adiabatic lapse rate of 10° C per kilometer. In this situation, a dry air parcel will rise, become colder than its environment, and sink back down. Because the parcel returns to the altitude where it started, it is a stable situation. However, a saturated air parcel will become warmer and warmer than its environment as it rises and keeps on rising. In such unstable situations, tall clouds can form, especially thunderstorms.

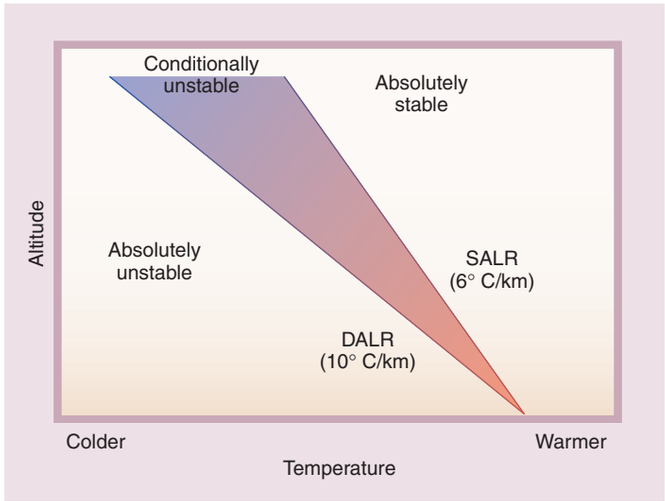


Figure 4.15
An environment can be described in three different ways in terms of lapse rates and stability: (1) Absolutely unstable, when the environmental lapse rate is greater than the dry adiabatic lapse rate (DALR; left); (2) conditionally unstable, when the environmental lapse rate is in between the DALR and the saturated adiabatic lapse rate (SALR; middle wedge); and (3) absolutely stable, when the environmental lapse rate is less than the SALR (right). In a conditionally unstable environment, saturated parcels are unstable and will keep on rising, leading to the development of tall clouds.

Figure 4.15 depicts our expanded understanding of static stability, including conditional instability. If you prefer words to figures, Table 4.2 summarizes the same information. Either way, the bottom line is that condensed moisture improves the ability of air to rise, form clouds, and cause "bad weather." In Chapter 11, we will explore how severe thunderstorms can be understood and predicted using these same concepts of saturation and stability.

TABLE 4.2 Atmospheric Stability Summary

Environmental Lapse Rate Is...	Environment Is...	Means What?
Less than saturated adiabatic lapse rate	Absolutely stable	No parcels keep rising
Greater than dry adiabatic lapse rate	Absolutely unstable	All parcels keep rising
Less than dry adiabatic lapse rate *and* greater than saturated adiabatic lapse rate	Conditionally unstable	Only saturated parcels keep rising

Cloud Classification

Earlier, we learned that fogs are named for the process that caused the air to become saturated. In 1803 British pharmacist and chemist Luke Howard devised a different kind of classification system for clouds above the ground. It has proved so successful that meteorologists have used Howard's system ever since, with minor modifications. According to his system, clouds are given Latin names corresponding to their appearance—layered or

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convective—and their altitude. Clouds are also categorized based on whether or not they are precipitating.

Layered clouds are much wider than they are tall. They generally have flat bases and tops and can extend from horizon to horizon. The Latin word *stratus* describes the layered cloud category, just as "stratosphere" describes a layered region of the atmosphere. Stratus-type clouds form in relatively stable air.

Convective clouds are as tall, or taller, than they are wide. These clouds look lumpy and piled up, like a cauliflower. Convective cloud types are indicated by the root word *cumulo*, which means "heap" in Latin. Convective clouds may become very tall and are rounded on top. They generally form in unstable air.

Clouds can also be classified by their altitude and their ability to create precipitation. The root word *cirro* (meaning "curl") describes a high cloud that is usually composed of wispy ice crystals. The Latin word *alto* ("high") is used to indicate a cloud in the middle of the troposphere that is below the high cirro-type clouds (just as altos in a choir sing lower notes than sopranos, but higher notes than basses). The prefix or suffix *nimbus* ("rain") denotes a cloud that is causing precipitation.

Using the combination of appearance, altitude, and ability to make precipitation, a wide range of cloud types can be identified. Table 4.3 classifies ten common cloud types and their typical heights above the ground, and Figure 4.16 depicts them pictorially. Now we will examine each cloud type, from the ground up.

MeteorologyNow™ Click the *Meteorology Interactive* "Name That Cloud" to quiz yourself on the cloud types discussed in this section.

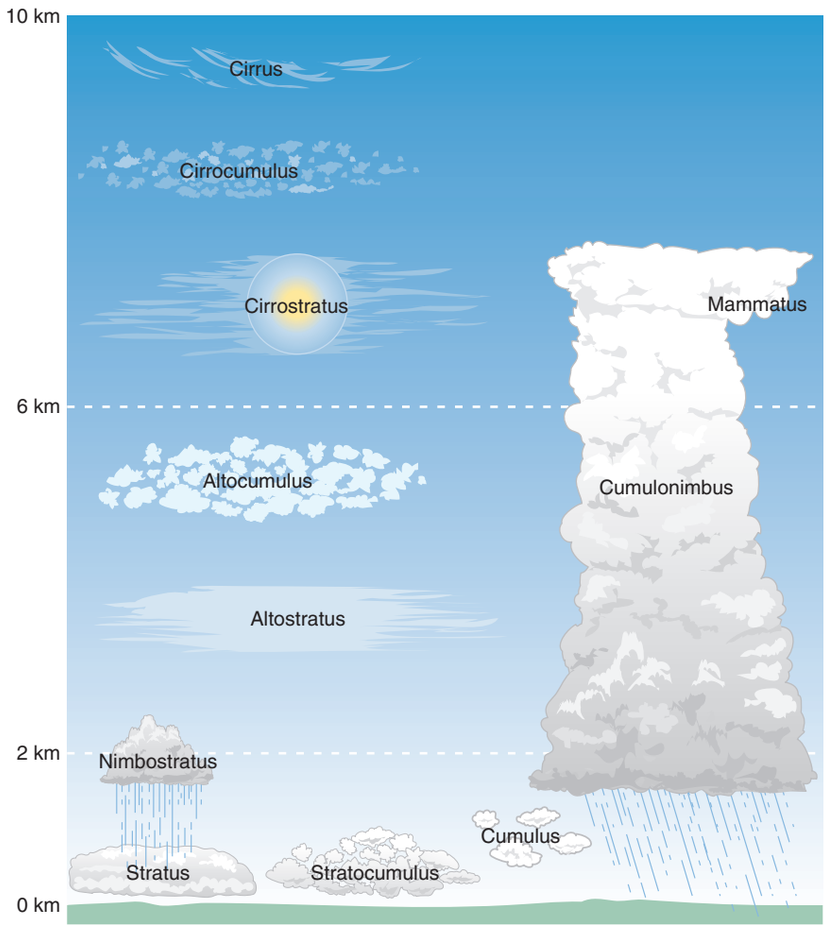


Figure 4.16
The major cloud types arranged by their typical altitude.

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TABLE 4.3 Common Cloud Types

				Typical Altitudes	
Cloud Type	**Layered Cloud**	**Convective Cloud**	**Mixed/Neither**	**Feet**	**Kilometers**
High	Cirrostratus	Cirrocumulus	Cirrus	20,000–40,000	6–12
Middle	Altostratus	Altocumulus		6500–20,000	2–6
Precipitating	Nimbostratus			Surface–10,000	0–3
		Cumulonimbus		Surface–50,000	0–15
Low	Stratus	Cumulus	Stratocumulus	Surface–6500	0–2

Low Clouds

Stratus (St)

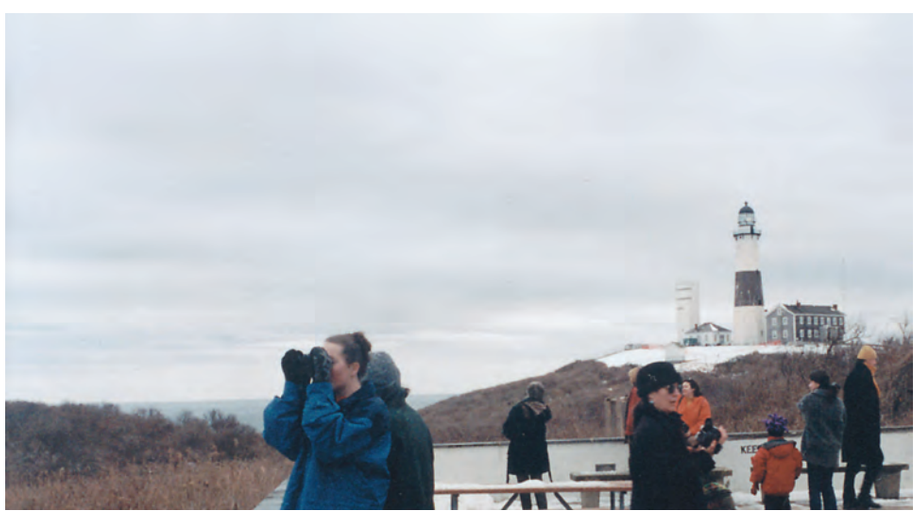
Stratus clouds, abbreviated St, are fog that hovers just above (rather than on) the ground. They are fuzzy and featureless in appearance (Figure 4.17). From the ground these clouds appear light to dark gray in color and cover the sky. They are common along coastlines and in valleys. Early morning fogs may lift and form a stratus cloud. Stratus clouds may also originate when moist, cold air is advected at low altitudes over a region. No precipitation normally occurs with stratus clouds, although a fine mist or drizzle sometimes is visible on car windshields during thick stratus.

Stratocumulus (Sc)

Stratocumulus (Sc) clouds are low-lying clouds that cover the sky and appear white to gray in color (Figure 4.18). They are a combination of layered and convective cloud types. This is because stratocumulus clouds often occur in a shallow layer of unstable air near the surface that is overlain by stable air.

Unlike featureless stratus clouds, stratocumulus clouds often appear in rows or patches. You can distinguish stratocumulus from stratus by looking for more variations in color and a lumpier appearance.

Stratocumulus are common in certain regions, such as coastlines and in valleys. Marine stratocumulus layers are very persistent off the California and South American coastlines. In those regions, moist air flows over cooler waters and becomes saturated. Stratocumulus clouds are also associated with fronts. When accompanying a large weather



Lil Ackerman

Figure 4.17
Stratus (St) are low-altitude layered clouds.

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Jeffrey Kay

Figure 4.18
Stratocumulus (Sc) are low-lying clouds with both layered and convective aspects. Stratocumulus are distinguished from stratus clouds by variations in color across the sky.

system, stratocumulus clouds are often the last clouds to appear before the skies clear completely.

Precipitation does not occur with stratocumulus. However, if the unstable surface air grows deeper (e.g., as a result of daytime heating), the stratocumulus can grow taller and develop into convective clouds that produce rain or snow.

Cumulus (Cu)

Cumulus (Cu) clouds generally have well-defined, flat bases and intricately contoured domed tops resembling cauliflower. The edges of the cloud are distinct. The bases are generally dark gray and the sunlit sides are bright white.

These clouds form whenever fairly humid air rises, usually by convection. The height of the bottom of the cloud (the *cloud base)* is related to the temperature and the dew point of the rising air. Cumulus clouds in the dry southwest United States generally have much higher bases than those in the Southeast. Cumulus clouds may also form over mountains or large hills if the air is unstable. These orographically forced clouds appear stationary, although they continually form and dissipate.

The two basic forms of cumulus clouds are *fair-weather cumulus* and *cumulus congestus*. Fair-weather cumulus clouds symbolize pleasant weather conditions all over the world (Figure 4.19). They have a height similar to their width. These clouds are common in summer when solar heating of the surface triggers convection. During autumn and winter, cumulus clouds often form in cold air over large open lakes that are still warm. Fair-weather cumulus are not deep enough to cause rain, although some may grow into large storms.

Cumulus congestus or towering cumulus are tall relative to their width. For these clouds to form, the atmosphere must have a deep unstable layer, deeper than is required for the formation of the fair-weather cumulus. These towering clouds are common in summer and may have light rain falling from them. In regions like Florida, *cumulus congestus* may produce heavy rains for a few minutes. When *cumulus congestus* form in the morning it is a good indicator that storms may form later in the day. If the cloud tops appear fuzzy, ice is forming and the cloud may be developing into a cumulonimbus.

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Steve Ackerman

Figure 4.19
Cumulus (Cu) clouds are often observed on summer days.

Precipitating Clouds

Nimbostratus (Ns)

Nimbostratus (Ns) are deep clouds that bring precipitation and appear dark gray to pale blue in color (Figure 4.20). The cloud base is difficult to see because precipitation is falling from the cloud. For this reason, nimbostratus sometimes look similar to stratus, stratocumulus, or altostratus clouds. Nimbostratus clouds often precede warm fronts.



Courtesy Elise Collet

Figure 4.20
Nimbostratus (Ns) are deep layered clouds that bring precipitation and appear dark gray to pale blue in color.

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David W. Martin, SSEC, University of Wisconsin-Madison

Figure 4.21
Cumulonimbus (Cb) over Lake Wingra, Madison, Wisconsin.

The precipitation that falls from nimbostratus clouds is usually continuous and light to moderate in intensity. More episodic and intense precipitation is associated with cumulonimbus clouds.

Cumulonimbus (Cb)

Cumulonimbus (Cb) are thunderstorm clouds. They extend upward to high altitudes, often to the tropopause and sometimes into the lower stratosphere. Cumulonimbus clouds produce large amounts of precipitation, severe weather, and even tornadoes (Figure 4.21).

A distinguishing feature of cumulonimbus is the flattened **anvil** shape of the top of the cloud. The anvil develops when the updraft slows down and spreads outward horizontally as it encounters the very stable air in the stratosphere. Underneath the anvil, sinking air may create pouches called **mammatus** (Figure 4.22). Although mammatus clouds are not severe weather, they can form under the anvils of strong thunderstorms.

Cumulonimbus clouds develop in unstable, moist atmospheres and are fairly common in the United States in spring and summer. They often occur ahead of cold fronts. In summer they can form over mountains because of orographic lifting in combination with solar heating. Cumulonimbus clouds can be isolated or organized in groups. When cumulonimbus clouds develop into an organized system the chance of severe weather often increases, as we will see in Chapter 11.

Middle Clouds

Altostratus (As)

Altostratus (As) are layered clouds made up mostly of liquid water droplets and are gray to pale blue in appearance (Figure 4.23). Altostratus form when the middle layers of the

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Greg Thomson

Figure 4.22
Pouchy mammatus clouds (top) *sometimes form on the underside of cumulonimbus anvils.*

atmosphere are moist and slowly lifted. If the Sun appears through these clouds, it has a "watery" appearance, whereas stratus clouds normally obscure the Sun. Altostratus clouds are common ahead of a warm front, before the nimbostratus.

Altocumulus (Ac)

The appearance of **altocumulus** (Ac) clouds varies considerably. They can be thin or thick, white or gray, and organized in lines or randomly distributed. They occur in the middle levels of the atmosphere when the air is moist, not too stable, and is being lifted. They are similar in appearance to stratocumulus, although with a higher cloud base (Figure 4.24).

You can distinguish between various types of cumulus clouds using the "fist-thumb-pinkytip" rule. Because of distance, clouds that are higher up appear smaller to your eye than those closer to the ground. If you extend your arm on a line from your eye to the cloud, cumulus clouds are generally about as big as your fist. Altocumulus clouds, in contrast, are only as big as your thumb. If the lumps of cumulus are even smaller, as small as

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K. G. Diem

Figure 4.23
Altostratus clouds (As) are layered clouds that exist in the middle layers of the troposphere and give the Sun or Moon a "watery" appearance.

the tip of your little finger, then the cloud is probably cirrocumulus (described in the next section).

Altocumulus clouds often appear ahead of a warm front, prior to altostratus. If other cloud types accompany altocumulus, a storm is probably approaching. Also, if the altocumulus rises up in little castle-like turrets called *castellanus* (Figure 4.25), it signals that the air aloft is unstable.

High Clouds

Cirrocumulus (Cc)

Cirrocumulus (Cc) clouds are thin, white clouds that often appear in ripples arranged in a regular formation (Figure 4.26). The smaller size of the individual cumulus lumps in cirrocumulus clouds distinguish this cloud type from altocumulus. Cirrocumulus clouds are



Lil Ackerman

Figure 4.24
Altocumulus (Ac) occur in the middle levels of the atmosphere when the air is moist.

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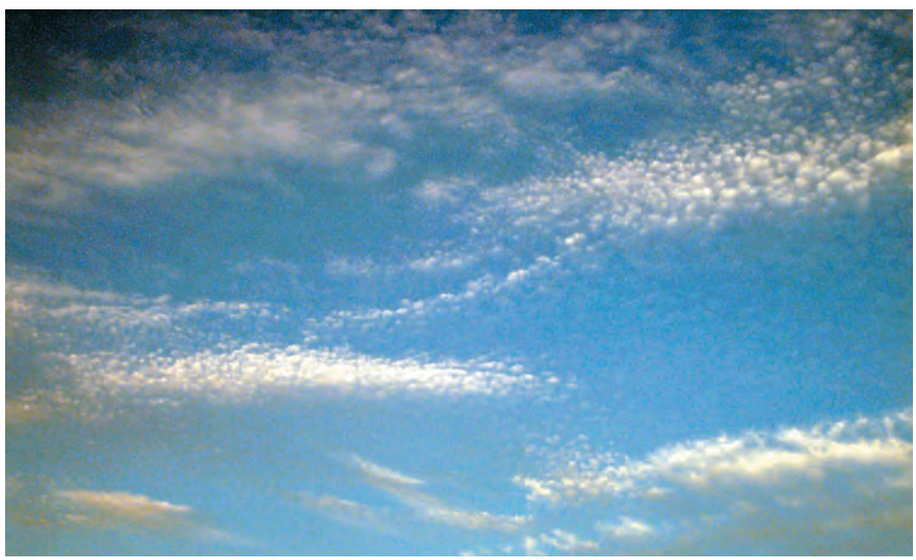
Don Lloyd

Figure 4.25
Altocumulus castellanus clouds grow into turret-like tufts and indicate unstable air in the middle troposphere.

composed of ice crystals and occur high in the atmosphere in regions that are relatively moist and unstable.

Although these clouds occur year-round, they are not very common and are usually present with other cloud types. Their tiny, delicately shaped features make cirrocumulus among the most beautiful of clouds. A "mackerel sky" is one that contains cirrocumulus clouds (or small altocumulus clouds) in a pattern that resembles fish scales.

Cirrocumulus clouds appear in association with large precipitation-causing weather systems, especially warm fronts. This cloud type usually follows cirrus and precedes



K. G. Diem

Figure 4.26
Cirrocumulus (Cc) are thin, white clouds that appear high in the troposphere.

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© J. Deguara/Storm Chaser

Figure 4.27
Cirrostratus (Cs) are layered clouds that are sometimes observed in connection with optical effects, such as halos and sundogs.

altocumulus as a precipitation-causing warm front approaches. For this reason, the saying "Mackerel sky, not three days dry" became a popular piece of weather folklore in the days before modern weather forecasting.



Greg Thomson

Figure 4.28
Cirrus (Ci) are wispy, fibrous, white clouds that are composed of ice crystals.

Cirrostratus (Cs)

Cirrostratus (Cs) clouds can cover part or all of the sky. They are uniform in appearance and can be thin or thick and white or light gray in color (Figure 4.27). Sometimes cirrostratus clouds are almost invisible and the Sun shines through easily, unlike the "watery sky" of altostratus. They occur high in the atmosphere and are composed of ice crystals.

Cirrostratus clouds are common during winter in association with large-scale weather systems. If the cirrostratus cloud thickens into altostratus, an approaching storm is indicated. They may also appear far out in advance of a tropical or subtropical weather disturbance.

Cirrostratus clouds are most famous for the optical effects that occur when the Sun or Moon shine through them. Halos, bright arcs, and brilliant spots form when light passes through the ice crystals composing the cirrostratus. We examine these optical effects in the next chapter.

Cirrus (Ci)

Cirrus (Ci) are wispy, fibrous, white clouds that are made of ice crystals. They often occur as wisps here and there across the sky that are aligned in the same direction as the upper-level winds (Figure 4.28). They are a very common cloud type associated with all weather systems, including fair-weather high-pressure areas. Cirrus clouds precede warm fronts and accompany jet streams. Mountains can also generate cirrus clouds when air is forced over high peaks. When isolated cirrus occur, they do not indicate approaching bad weather. *Mares' tails* are cirrus clouds that are long and flowing, like a horse's tail.

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There are many other types of clouds, including some that are associated with specific weather and climate patterns such as the ozone hole and mountain wind circulations. We will examine those clouds in later chapters when we discuss the phenomena associated with them.

Clouds and the Greenhouse Effect

Before we delve into the details of cloud composition, let's discuss their crucial role in the global warming debate. As we learned in Chapter 2, greenhouse gases such as water vapor and carbon dioxide warm the atmosphere by absorbing the longwave radiation emitted from the surface. Water vapor is an important greenhouse gas because it absorbs longwave energy effectively. Absorption of that energy warms the atmosphere. Increases in greenhouse gases over time can potentially result in a climate change because the atmosphere becomes more effective at absorbing longwave energy emitted by the surface.

If the atmosphere warms, initially the relative humidity should decrease. Evaporation depends on relative humidity. With a lower relative humidity, more evaporation occurs, which adds more water molecules to the atmosphere and enhances the greenhouse warming. Increases in the temperature of the atmosphere would affect the dynamics of weather and climate.

Greenhouse gases are not the whole story, however. Clouds have a large impact on the energy gains of the atmosphere. Clouds reflect solar energy into space, away from the air beneath them, and clouds reduce the amount of solar radiation reaching the surface. Because of this, clouds tend to cool the Earth (Figure 4.29). The thicker the cloud, the more energy reflected back to space and the less solar energy available to warm the surface and atmosphere below the cloud. By reflecting solar energy back to space, clouds tend to cool the planet.

Clouds also have a warming effect on atmosphere below them because they are very good emitters and absorbers of terrestrial radiation. Clouds block the emission of longwave radiation to space and inhibit the ability of the planet to emit its absorbed solar energy to space in the form of longwave radiation (see Figure 4.29). Thus, in the longwave, clouds act to warm the planet, much like the greenhouse gases do.

To complicate matters, the altitude of a given cloud is important in determining how much it warms the planet. Cirrus are cold clouds. Thick cirrus clouds emit very little energy because of their cold temperature, according to the Stefan-Boltzmann Law from Chapter 2. At the same time, cirrus clouds are effective at absorbing the surface-emitted heat. Thus, with respect to longwave radiation losses to space, cirrus clouds tend to warm the planet.

Stratus clouds also warm the planet but not as much as cirrus. This is because stratus clouds are low in the atmosphere and have temperatures that are closer to surface temperatures than cirrus clouds. Stratus clouds absorb radiation emitted by the surface, but, because these low clouds have a similar temperature as the surface, they emit similar amounts of terrestrial radiation to space as the surface.

To complicate matters still further, a cloud's effectiveness at reflecting sunlight is related to how large the cloud droplets or cloud ice crystals are. We will investigate the reasoning behind this in Chapter 5, but it is easily demonstrated with ice in a familiar form. If you look at a glass filled with crushed ice next to a glass filled with ice cubes, you can see that the crushed ice is brighter (whiter) than the glass of ice cubes. The crushed ice particles are smaller than the cubes. Similarly, clouds consisting of small droplets are brighter than clouds consisting of large particles. Clouds composed of small particles therefore have a higher albedo and reflect more solar radiation back into space, causing more cooling than clouds with large particles.

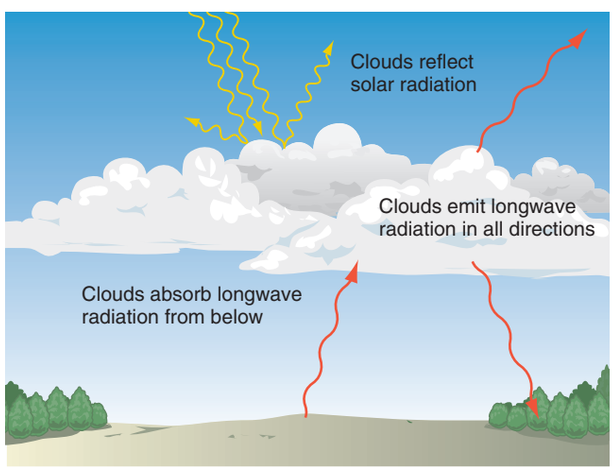


Figure 4.29
In the solar spectrum clouds tend to cool Earth; in the longwave spectrum they tend to warm the planet.

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In summary, clouds can act to cool or warm the planet, depending on how much of the Earth they cover, how thick they are, how high they are, and how big the cloud particles are. Measurements by NASA indicate that on average, the reflection of sunlight by clouds more than compensates for the clouds' greenhouse warming. Thus, today's distribution of clouds tends to cool the planet.

This may not always be the case, however. As the atmosphere warms, the distribution of cloud amount, cloud altitude, and cloud thickness all may change. We do not know what the effect of clouds will be on the surface temperatures if the global climate changes. Clouds could dampen any greenhouse warming by increasing cloud cover or decreasing cloud altitude. On the other hand, clouds could increase a warming if the cloud cover decreases or cloud altitude increases. Climate is so sensitive to how clouds might change that an accurate prediction of climate change hinges on correctly predicting the fine details of cloud formation and composition.

Cloud Composition

Every cloud has a unique composition. The composition of a cloud includes the phase of the water in it, the number and size of particles, and the shape of any ice particles, if they are present.

Water-bearing clouds differ in composition over land versus over the oceans. There are more cloud condensation nuclei over land than over the oceans. For this reason, continental clouds tend to have a greater number of water droplets than maritime clouds. Clouds over land have approximately 500 million to 1 billion cloud droplets per cubic meter of air; maritime clouds have about one tenth as many. Because the water content of maritime and continental clouds are similar, however, the drops in continental clouds are usually smaller and more numerous than the maritime counterparts. Maritime clouds have large, soluble, heterogeneous nuclei that favor the formation of large droplets.

Ice-containing clouds vary greatly in terms of the number, size, and shape of the ice crystals in them. The size and shape of a crystal is called its **crystal habit.** Temperature determines the particular crystal habit of ice. Figure 4.30 shows the four basic shapes of ice crystals, each of which occur preferentially in the following temperature ranges: the **hexagonal plate** (0° to −5° C; −10° to −12° C; −16° to −25° C), the **needle** (−5° to −10° C), the **column** (−5° to −10° C; −25° to −50° C), and the **dendrite** (−12° to −16° C). The dendrites are hexagonal with elongated branches, or fingers, of ice. They most closely resemble what we think of as snowflakes. We will soon learn that this is because ice crystals grow fastest around −15° C, the range in which dendrite formation is preferred.

MeteorologyNow™ Click the *Atmospheric Exploration* "Growing a Snowflake" to make your own beautiful and complex ice crystals.

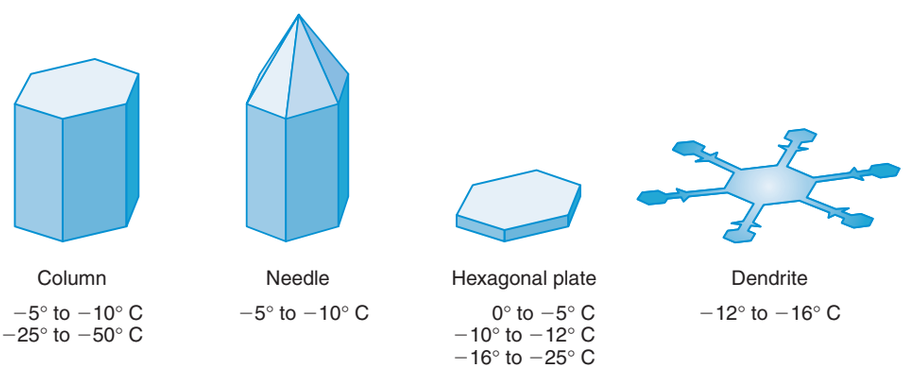


Figure 4.30
The four basic ice crystal habits are column, needle, hexagonal plate, and dendrite. The shape in which an ice particle grows depends on the temperature of its environment. Try this out for yourself by using the "Growing a Snowflake" learning applet.

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Precipitation

Precipitation is any liquid or solid water particle that falls from the atmosphere and reaches the ground. Precipitation can be long-lasting and steady, or it may fall as a brief and intense shower. Because precipitation is formed from water vapor, it removes water vapor from the atmosphere. Rain, snow, sleet, freezing rain, and hail are all forms of precipitation.

Dew and frost also remove water vapor through condensation or deposition onto surfaces on the ground. Dew and frost are not precipitation, however, because they do not fall from a cloud under the force of gravity. In precipitation, water vapor condenses onto a particle that eventually grows large enough to fall out of a cloud and to the surface. After discussing how particles grow into precipitation, we examine the main precipitation types.

Precipitation Growth in Warm Clouds

Rainmaking, natural or artificial (Box 4.2), is not easy. Cloud particles are usually 10 microns (μm) in size. (For comparison, the period at the end of this sentence is about 500 μm in diameter.) Small raindrops are usually 1000 μm in diameter. About 1 million cloud droplets have to combine to form a single raindrop. How does this happen? Just as with cloud particle growth, the formation of precipitation is complicated by the different properties of water at different temperatures.

The simple, but incorrect, explanation for precipitation is growth by condensation. Why can't a cloud droplet grow into a raindrop by condensing water onto its surface?

BOX 4.2 Controlling the Weather

It's an age-old question: Can humans control the weather? In the past, people rang bells or fired cannons to prevent lightning or cause rain—producing sound and fury, but nothing in the way of success.

The scientific era of weather modification began in the 1940s and 1950s with the advent of cloud-seeding experiments. In cloud seeding, airplanes drop particles of dry ice or silver iodide into cold clouds. These particles act as extremely effective ice nuclei, potentially increasing the amount of rainfall or snowfall. However, progress in cloud seeding has been slow. Today, scientists agree that the seeding of orographic clouds can enhance precipitation by a modest 10% on a seasonal basis. How this actually happens is still poorly understood.

Farmers have long sought for a way to suppress hail. One hailstorm can destroy a year's crops in a few minutes. Cloud seeding during the early stages of cumulonimbus development can reduce hail damage by keeping hailstone sizes small, but the results so far have been mixed. Research on hail suppression continues.

Another active area of weather modification is fog dispersal. Fog can shut down an airport for hours, causing delays and cost overruns. Like clouds, fog can be seeded with materials that cause the water droplets in fog to turn into ice, which cleanses the air of fog. This is now a routine practice at many airports worldwide. However, it is impractical over large regions.

At one time or another you may have wondered, "Why can't meteorologists make tornadoes and hurricanes go away?" Perhaps a nuclear explosion or, less violently, cloud seeding could disrupt the circulation of a tornado or a hurricane. Alternatively, oceans could be covered with water-impermeable chemicals, cutting off a hurricane's fuel source.

Severe weather is here to stay, however. Hurricane seeding in the 1950s and 1960s produced few firm results. Since then, research on this subject has been curtailed. Tornadoes are even less well understood than hurricanes, and there is currently no anti-tornado research in progress.

Legal and ethical dilemmas arise whenever severe weather modification is proposed. The atmosphere cannot be put into a test tube, and so experiments must be conducted on actual storms. However, what if a cloud-seeded hurricane hits New York City instead of moving out to sea harmlessly? Who is to blame? How would the environment respond if the Gulf of Mexico were covered with a film of chemicals? Obviously, the consequences of exploding nuclear bombs in storms would be horrific. The difficulty of weather modification, combined with the risks and dilemmas associated with it, have kept work on this subject to a minimum.

As we will see throughout this text, humans have changed the atmosphere in many ways. These changes usually come about as unintended by-products of modern civilization. So far, humans have been more effective at modifying weather and climate by mistake than by design.

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Because this process does not work fast enough, except at the smallest drop sizes. It would take several *days* for a full-sized raindrop to form by condensational growth. Precipitation forms much more quickly than that. There must be another mechanism for the relatively rapid growth of precipitation-sized particles.

Collision-Coalescence

One process that could produce a larger drop quickly would be to combine several smaller particles. To do this the cloud particles have to bump into each other and merge together, or *coalesce*. This is called the **collision-coalescence** process.

To explain how this process works, let's consider water droplets in an updraft. Water droplets in clouds with different sizes move at different speeds, as gravity and vertical motions act on them. The difference in speed increases the chance of collisions, just as the combination of fast trucks and slow cars increases the chance of collisions on a highway.

Just because two droplets approach one another does not guarantee they will collide. A large droplet in motion creates an air current around it that can be strong enough to force tiny droplets to flow around it. (For the same reason, it is rare for cars on a highway to hit flying birds because the air current around the car carries the bird around, not into, the car.) This reduces the probability that droplets will collide, although initially they may be headed straight for one another (Figure 4.31). Once a drop grows to a size where the downward force of gravity exceeds the updraft force of air currents, it falls downward through the cloud. As the drop falls through the cloud, it sweeps up smaller droplets in its path, collecting them and growing by collision-coalescence.

Precipitation Growth in Cold Clouds

The process of combining particles through collision-coalescence is an important mechanism for forming precipitation in clouds composed solely of liquid water droplets. Therefore, it is most effective in "warm" clouds in the tropics. Outside of the tropics, clouds contain ice particles, even in the summertime. How does precipitation form in these cold clouds?

Accretion and Aggregation

Collision also helps create precipitation in cold clouds. When an ice crystal falls through a cloud it may collide with and collect supercooled water drops. This process of ice crystal growth by sweeping up supercooled water drops is called **accretion.** When ice crystals collide with supercooled drops, the drops freeze almost instantly. Accretion thus provides a mechanism for the particle to grow quickly.

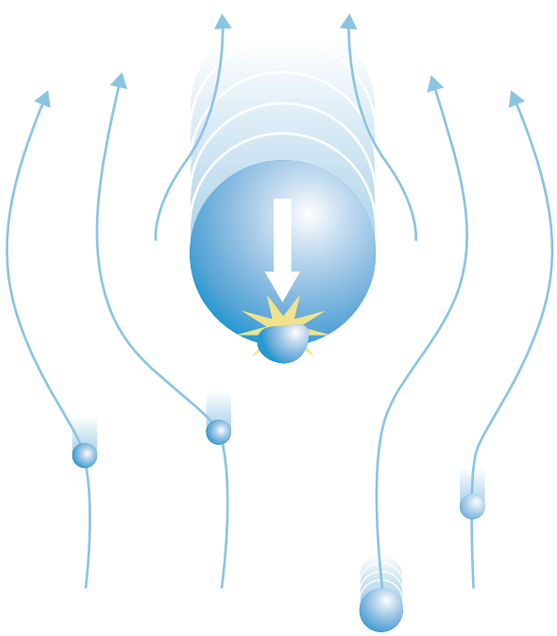


Figure 4.31
Large water drops fall faster than smaller ones. Because of the different fall speeds, water drops sometimes collide and coalesce. Very small droplets may flow around the larger drops and avoid colliding with them.

An ice particle produced by the accretion process that has a size between 1 millimeter and 5 millimeters (0.04 to 0.2 inches) and no discernible crystal habit is called **graupel** (pl. *graupeln*). On collision and freezing, the supercooled water often traps air bubbles. Because of this trapped air, the density of a graupel is low and it can easily be crushed, unlike a hailstone.

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Aggregation is the process by which ice crystals collide and form a single larger ice particle (Figure 4.32). The probability that two crystals will stick together depends on the shape of the crystals. If two dendrites collide, it is likely that their branches will become entangled and the two crystals will stick together. When two plates collide there is a good chance that they will simply bounce off one another.

Temperature also plays a role in aggregation. If the temperature of one crystal is slightly above freezing, it may be encased in a thin film of liquid water. If this particle collides with another crystal, the thin film of water may freeze at the point of contact and bond the two particles into one. (This is why you should never lick a cold metal flag pole!)

A **snowflake** is an individual ice crystal or an aggregate of ice crystals. Snowfalls do not consist of single crystals. More commonly, they are composed of flakes that are aggregates of ice crystals. Snowflakes composed of aggregates can sometimes reach 3 or 4 inches in size.

In both warm and cold clouds, how big a droplet or crystal grows by collision processes depends on how long it stays in the cloud. The longer a particle is in the cloud, the more particles it can collect and the larger it grows. The strength of the updrafts and the thickness of the cloud determine how long it stays in the cloud. This is why only tall clouds with strong updrafts, such as cumulonimbus clouds, produce large precipitation particles. In contrast, stratus clouds are shallow in depth and have much weaker updrafts than cumulonimbus. Particles usually do not stay in a stratus cloud long, and large particles rarely form.

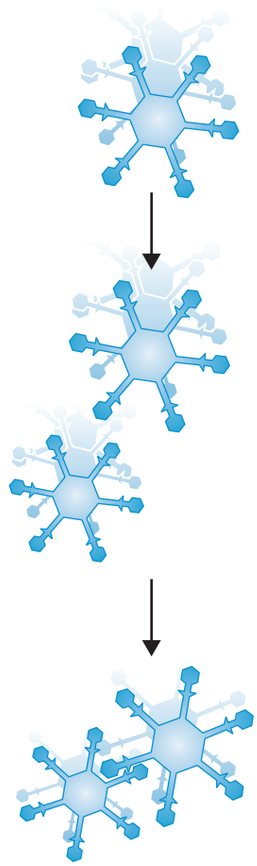


Figure 4.32
Ice crystals of different sizes or different shapes may collide and stick together, or aggregate.

The Bergeron-Wegener Process

The presence of ice in a cloud gives it a unique precipitation-making ability. To understand it, let's consider two sealed containers connected by a tube with a valve that can be opened (Figure 4.33). One container holds supercooled water at a temperature of −5° C (23° F), and the second contains ice at the same temperature.

The bonding forces in ice are much stronger than those in water, so fewer molecules have the energy to escape the ice than the number leaving the water. This means there will be fewer molecules of water in the vapor phase in the container with the ice than in the container that has the water—that is, the vapor pressure over water is greater than the vapor pressure over the ice. Because vapor pressure is proportional to the number of water molecules in the air, this also means that *the saturation vapor pressure is greater over water than over ice*. If the valve were opened between the containers, water molecules in the vapor would flow toward the region of lower vapor pressure over the ice.

Now consider an ice crystal surrounded by supercooled droplets (Figure 4.34). If the air is saturated (100% relative humidity) with respect to the water droplets, it is *super*saturated with respect to the ice crystals. Water vapor molecules will deposit onto the crystal, lowering the relative humidity of the air. In response, water molecules evaporate from the water droplet, supplying more water molecules to the air that then deposit onto the crystal.

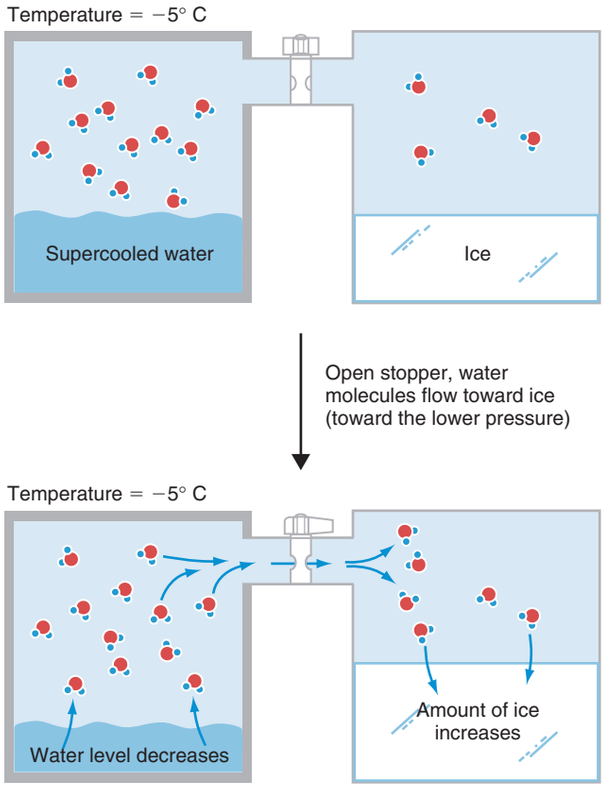
Put simply, *ice crystals grow at the expense of water droplets* in a cloud that has both. This ice crystal growth process is called the **Bergeron-Wegener process.** It was first proposed by meteorologist Alfred Wegener (who also famously proposed the theory of continental drift) in 1911 and explained more extensively by Tor Bergeron, a member of the renowned Bergen School of Norwegian meteorology. This process contributes to the rapid growth of ice crystals and, therefore, to the ability of a cloud to form precipitation.

This ice crystal growth process is fastest when the difference in the saturation vapor pressure between water and ice is large. As shown in Figure 4.35, these differences maximize when the air temperature is between −12° and −17° C (10.4° and 1.4° F). Based on what we have already learned about crystal shapes, this means that dendrites grow the fastest by the Bergeron-Wegener growth process. This explains why snowflakes look more like asterisks than plates, columns, or needles.

Precipitation Types

What happens when a particle falls out the base of a cloud? It is not officially precipitation until it reaches the ground. In some cases this never happens. For instance, if the

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Active Figure 4.33 MeteorologyNow™

At a given temperature, the saturation vapor pressure over ice is less than the saturation vapor pressure over water. As a result, water vapor is preferentially attracted to ice versus water.

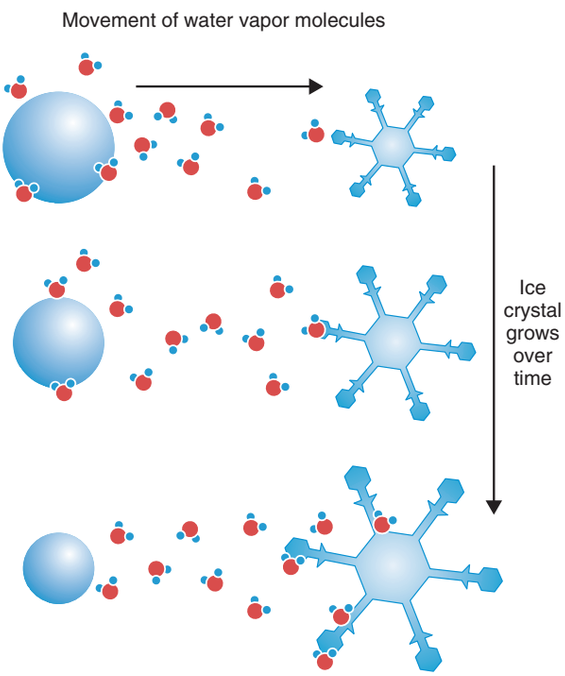


Figure 4.34

The larger saturation vapor pressure over a liquid water surface than over an ice surface causes the ice crystal to grow and the supercooled drops to evaporate.

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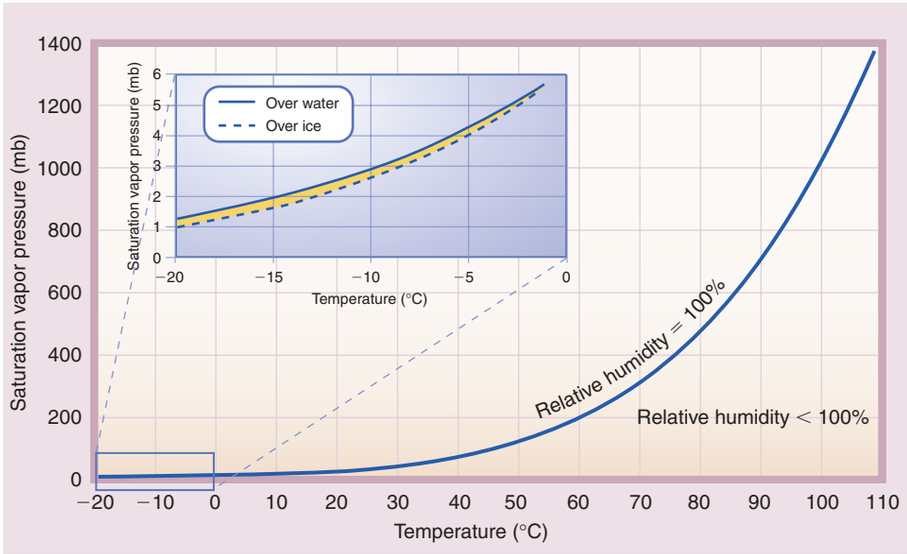


Figure 4.35
The differences between the saturation vapor pressure over ice and over water. The larger graph shows the saturation vapor pressure over water, which is the upper boundary of the observations in Figure 4.3. The inset compares the saturation vapor pressure over ice and over water. Because the line for water lies above the line representing ice, when the air is saturated (100% relative humidity) with respect to liquid water, it is supersaturated with respect to ice.

atmosphere below the cloud is very dry, the particle may evaporate in between the cloud base and the ground. **Virga** is rain that evaporates before reaching the surface.

Similarly, falling ice crystals may sublimate in dry air while being carried horizontally by the strong winds aloft. These **fallstreaks** (Figure 4.36) often produce visually striking patterns.

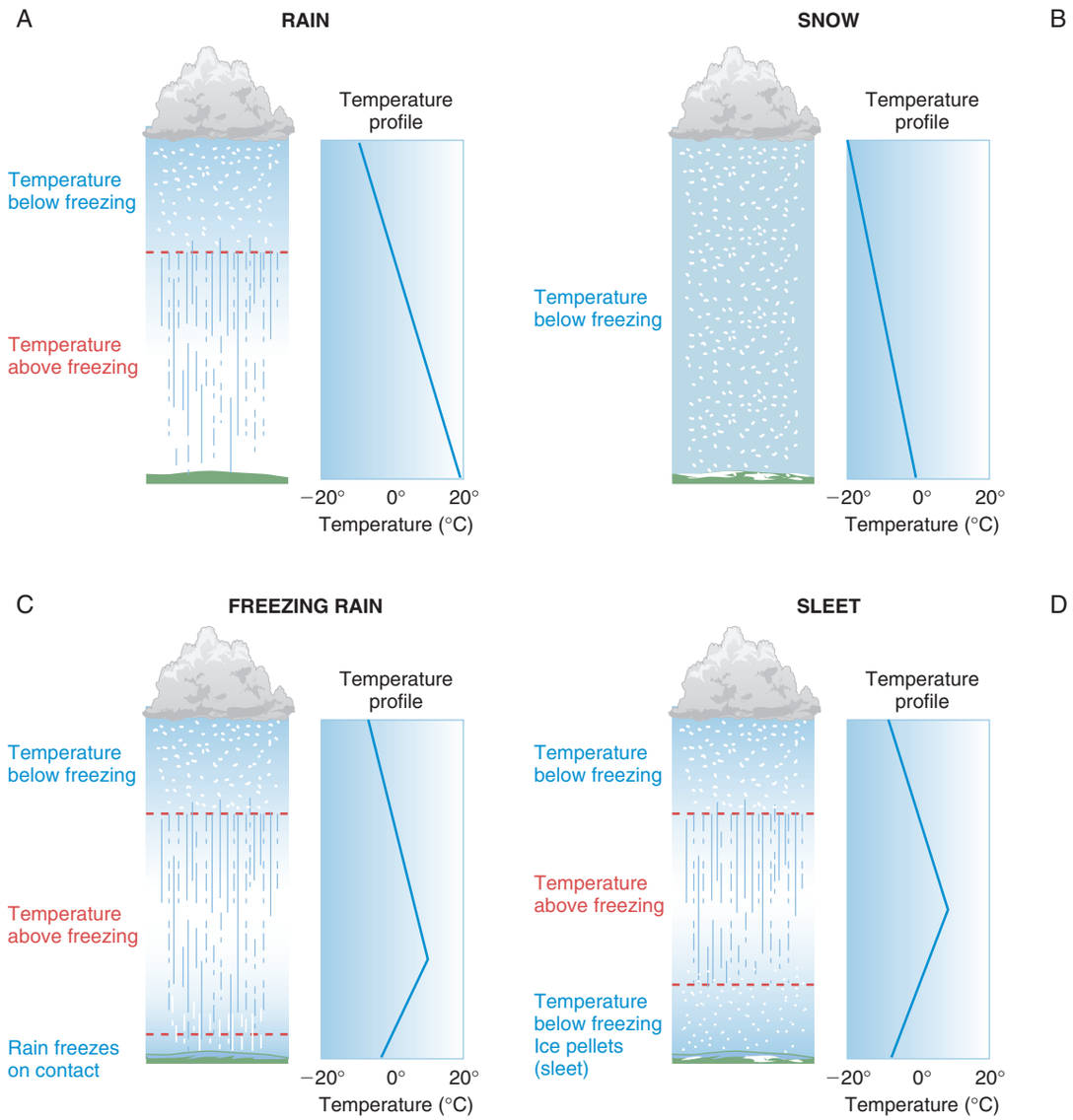
Meteorology Now™ Click the *Atmospheric Exploration* "Precipitation Type" to explore in detail the conditions that lead to various precipitation types.



Courtesy Andy Kilgore

Figure 4.36
Fallstreaks are wisps of ice particles that fall out of a cloud but evaporate before reaching the surface. Fallstreaks usually exhibit a hooked form produced by changes in wind speed with height.

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Active Figure 4.37 Meteorology Now™

*The vertical variation of temperature will determine if precipitation falls as rain (***A***), snow (***B***), freezing rain (***C***), or sleet (***D***). In this example, all the precipitation particles are initially ice crystals. Particles that fall into the melting layer become liquid drops. To form freezing rain or sleet a surface temperature inversion is required.*

When particles reach the surface as precipitation, they do so primarily as rain, snow, freezing rain, or sleet. Why are there so many different types of precipitation? Again, the ability of water to change phase is the key. In mid-latitude regions, precipitation usually begins as ice particles. Figure 4.37 shows the temperature conditions below the cloud base that lead to rain, snow, freezing rain, and sleet. The dashed line in Figure 4.37 represents the melting line, the altitude at which the temperatures are 0° C (32° F). The main difference between the different types of precipitation is whether an inversion exists near the ground—and if so, how thick the inversion is. Now, let's examine each precipitation type in detail, with reference to Figure 4.37.

Rain

If the temperature remains above 0° C (32° F) from the cloud base to the surface, precipitation particles melt into liquid droplets called **rain** (Figure 4.37, A). Raindrops are not

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teardrop-shaped; they are spheres flattened by the pressure of the wind as they fall. Raindrops are at least 500 μm (0.5 millimeters or 0.02 inches) in diameter. Precipitation drops smaller than this are collectively called **drizzle**, which is often associated with stratus clouds.

Average annual rainfall as estimated by satellite measurements is shown in Figure 4.38. The data in this figure mesh well with what we have learned about water in the atmosphere. For example, the warm tropical regions of the globe receive the most rainfall. This is partly a consequence of the fact that there is more water vapor in warm, moist air than in cooler air (see Figure 4.3). But why is there a narrow band of high rainfall near the equator? Also, why are the regions offshore of California and western South America so dry? To explain both of these enigmas we will need to understand global-scale wind patterns, the topic of Chapter 7.

Rain intensity is classified by the volume of rain that falls in an hour, according to the following scale:

Intensity	*Hourly Rainfall*
Light	0.25–2.5 millimeters (0.01–0.10 inches)
Moderate	2.5–7.6 millimeters (0.11–0.30 inches)
Heavy	More than 7.6 millimeters (0.30 inches)

However, extreme rainfall rates are possible. When this happens, severe flooding usually follows. On the Fourth of July in 1956, Unionville, Maryland, was drenched with a world-record 31.24 millimeters (1.23 inches) of rain in only 1 minute! More recently, on July 28, 1997, the campus of Colorado State University (CSU) at Fort Collins, Colorado received 134.6 millimeters (5.3 inches) of rain in less than 6 hours. The resulting flood inundated the CSU student center and library. All of the Fall 1997 semester textbooks were lost in the flooded campus bookstore, and 425,000 library books were soaked.

Snow

If the temperature underneath a cloud stays below freezing all the way to the ground, the snowflakes never melt and **snow** falls (Figure 4.37, *B*). As with rain, snow intensity is recorded in three categories. This is done according to volume of snowfall, or, because snow reflects and scatters light effectively, according to the reduction in visibility that results from the snowfall. The categories are as follows:

Intensity	*Hourly Snowfall*	*Visibility*
Light	Less than 0.5 centimeters (0.2 inches)	0.8 kilometers (0.5 miles) or more
Moderate	0.5–4 centimeters (0.2–1.5 inches)	0.4–0.8 kilometers (0.25–0.5 miles)
Heavy	Greater than 4 centimeters (1.5 inches)	Less than 0.4 kilometers (0.25 miles)

Water expands when it freezes, and snowflakes trap air between them when they clump together on the ground. For these reasons, an inch of snow and an inch of rain are not the same thing. A general rule is that 10 inches (25 centimeters) of new snow has the same water content as just 1 inch (2.5 centimeters) of rain. Cold, dry snows may be closer to a 20-to-1 ratio. However, wet snows with relatively warm temperatures can contain much more water, as much as 4 inches of water per 10 inches of snow.

These statistics can help us interpret the annual average snowfall across the United States, as shown in Figure 4.39. The northern Great Plains receive up to 5 feet of snow annually. When this thick snow cover melts, however, it is equivalent to just a few inches of rainfall. This is one reason why the Great Plains are dry. Elsewhere in Figure 4.39, it

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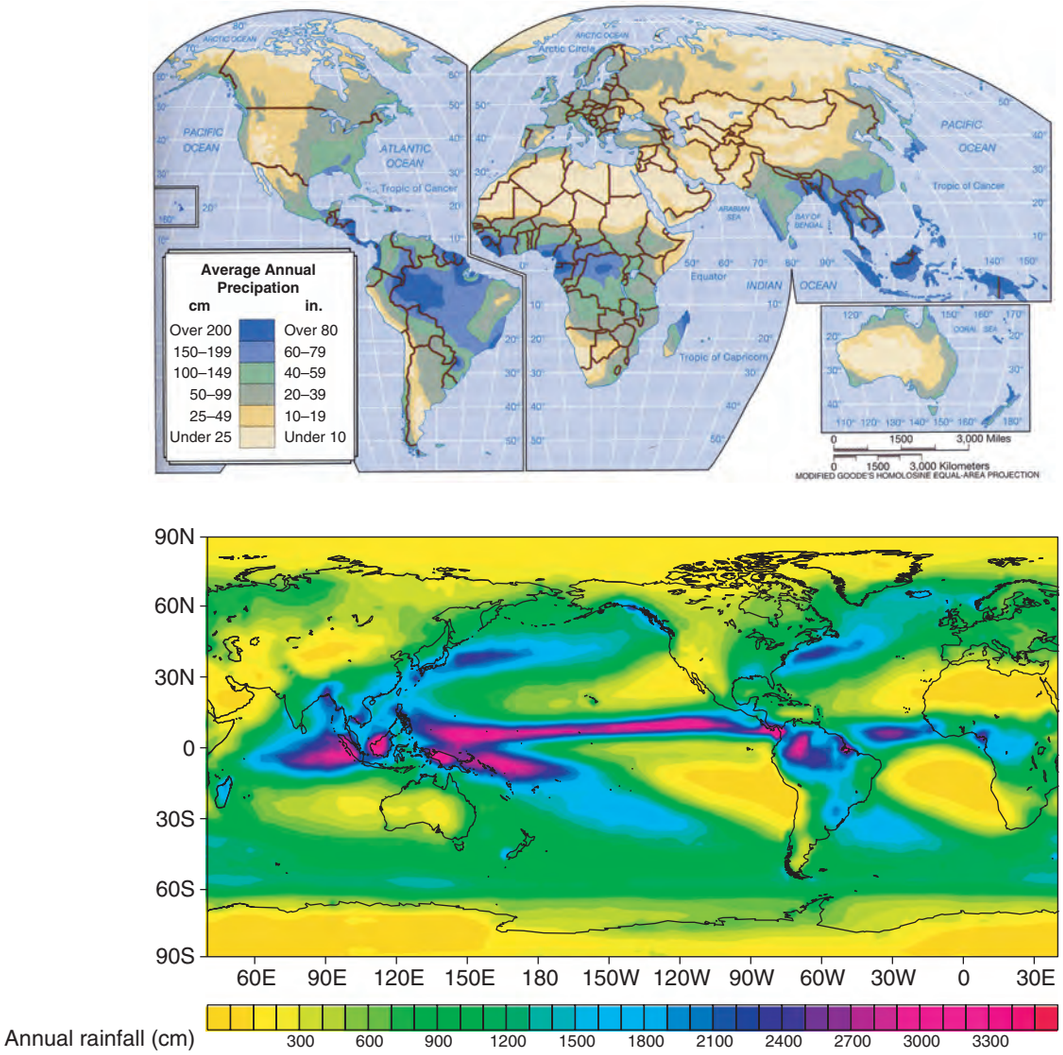


Figure 4.38

A: *Climatology of annual precipitation over the land regions of the world.* **B:** *Climatology of annual rainfall across the world, including the oceans, based on satellite observations.* *(http://www. jisao.washington.edu/legates_msu/index.new.html.)*

is apparent that the snowiest regions are in the Rocky Mountains, where cold temperatures and orographic lifting enhance snowfall.

Like rainfall, snowfall rates can be extreme. In Silver Lake, Colorado, on April 14 and 15, 1921, a whopping 195.6 centimeters (76 inches, or more than 6 feet) of snow fell in just 24 hours, a North American record.

Sleet and Freezing Rain

Ice storms occur when precipitation particles melt and then fall through a layer of cold air near the ground. The two precipitation types most common during ice storms are freezing rain and sleet.

Freezing rain forms when a thin layer of cold air near the surface causes melted precipitation to become supercooled (Figure 4.37, C). It then freezes on contact with exposed objects. Freezing rain covers everything in a sheet of ice, creating shimmering landscapes. However, even a little freezing rain causes treacherous road conditions and tree and power line damage. It is also responsible for aircraft icing, which is a cause of fatal plane accidents.

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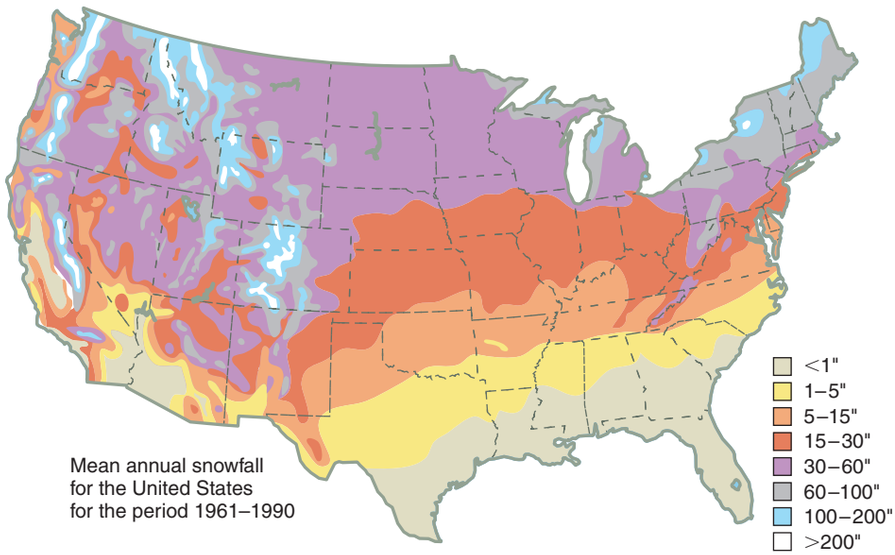


Figure 4.39
Climatology of the average annual snowfall across the lower 48 United States. Latitude and proximity to water strongly affect snowfall amounts over the eastern states, whereas elevation dominates in the western states.
(http://climate.atmos.colostate.edu/snowbooklet.shtml#Snmaps)

Figure 4.40 shows the number of hourly observations of freezing rain across the United States for the winter months of 1982 through 1990. Freezing rain is most common in the Northeast, the upper Midwest, the Columbia River valley, and along the mid-Atlantic foothills of the Appalachians. These are regions where cold air sinks or can be trapped at the surface (to be discussed in greater detail in Chapter 12). Meanwhile, warmer precipitating air moves above the cold air, creating an inversion. This thin layer of below-freezing air at the surface is a key ingredient for freezing rain.

Sleet consists of translucent balls of ice that are frozen raindrops. It occurs when the layer of subfreezing air at the surface is deep enough for the raindrop to freeze (Figure 4.37, *D*). Therefore, the difference between sleet formation and freezing rain formation is quite small, although the two precipitation types do not look at all alike. When sleet hits the surface, it bounces and does not coat objects with a sheet of ice, as freezing rain does. Instead it covers flat surfaces such as roads and driveways like millions of icy ball bearings.

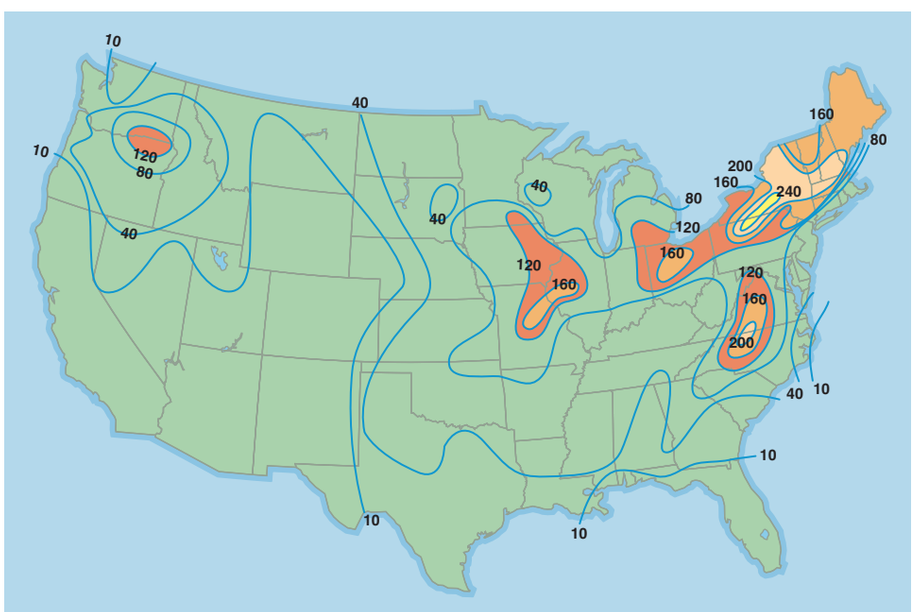


Figure 4.40
A 30-year climatology of freezing precipitation across the lower 48 United States, shown in terms of the number of hours of freezing rain per year.
(Adapted from http://rammb.cira.colostate.edu/visit/ptype/climofz.asp.)

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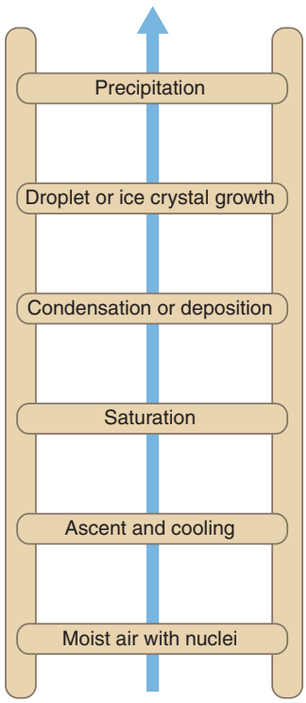


Figure 4.41
The steps by which clouds and precipitation are made, starting at the bottom and climbing to the top.

Keep in mind that the different scenarios in Figure 4.37 are idealized. Snow and sleet can occur at temperatures above 0° C when the air underneath the cloud is dry. In these cases, the precipitation particle partly evaporates as it falls. The evaporation cools the particle enough to keep it frozen all the way to the ground. Sleet seems particularly resistant to melting. One of us has seen sleet three separate times with surface temperatures at or above 10° C (50° F)! You can explore how and under what conditions different forms of precipitation occur using the Precipitation Types learning applet on the text's Web site.

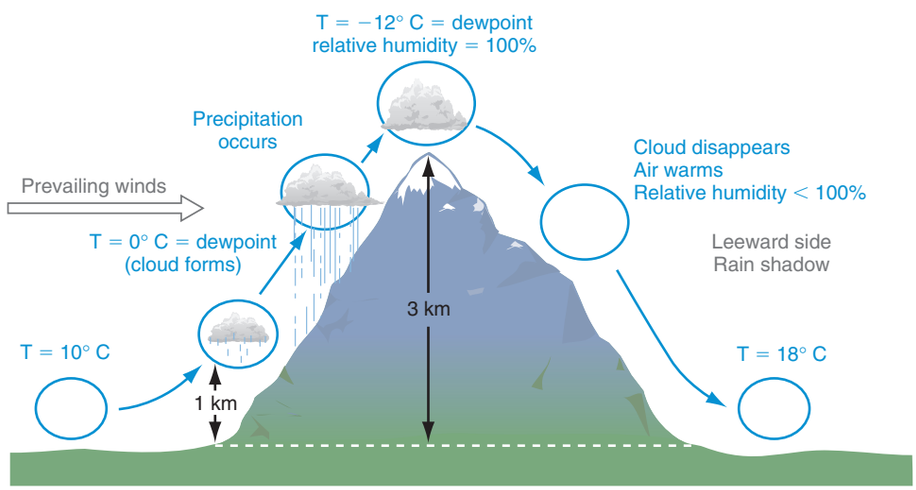
In addition to rain, snow, freezing rain, and sleet, other precipitation types exist. The most notable is **hail,** which is precipitation in the form of large balls or lumps of ice that look like sleet on steroids. The formation of hail is quite different than sleet formation, however. Hail develops in the complex air motions inside a towering cumulonimbus cloud. For that reason, we will discuss hail in connection with thunderstorms in Chapter 11. The "ladder" in Figure 4.41 summarizes the steps in the formation of the precipitation types we have discussed in this chapter.

Clouds, Lapse Rates, and Precipitation Near Mountains

In closing, let's investigate how concepts in this chapter integrate with concepts presented in previous chapters by analyzing airflow over a mountain (Figure 4.42). Our goal is to explain why semi-arid **rain shadow** regions exist downwind (the "lee side") of a large mountain range.

When air rises up a mountain, it expands and cools at the dry adiabatic lapse rate. As the temperature decreases, the relative humidity increases, and the temperature approaches the dew point temperature. Eventually, the temperature will equal the dew point temperature and a cloud will form, marking the cloud base.

If the air continues to rise, water vapor will continually condense to form cloud droplets. A phase change of water vapor to a liquid releases energy, warming the parcel through latent heating. This causes the parcel to cool more slowly, at the saturated adiabatic lapse rate. As the parcel continues to be orographically lifted, cloud droplets grow by condensation and by collision and coalescence. As the moist parcel continues to rise and cool, some of the liquid drops freeze. The cloud particles can then grow by the Bergeron-Wegener process, accretion, and aggregation.



Active Figure 4.42 Meteorology Now™
As moist air flows over a mountain, the temperature decreases, the air becomes saturated, clouds form, and precipitation falls on the upwind side. Sinking motions on the leeward side of the mountain cause the air to become warmer and drier than before. This generates a rain shadow downwind of the mountain.

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TABLE 4.4 Difference in Temperature, Cloud Cover, and Precipitation on the Windward and Leeward Sides of the Cascade Mountains in Washington State

	Windward (West) Side (Seattle–Tacoma)	*Leeward (East) Side (Yakima)*
Mean winter temperature	41° F	32° F
Mean summer temperature	64° F	68° F
Mean annual temperature range (warmest month's average minus the coldest month's average)	25° F	40° F
Number of mostly cloudy days per year	226	164
Average annual precipitation	37.2 inches	8.0 inches

Source: Western Regional Climate Center, http://www.wrcc.dri.edu/summary/lcd.html.

The precipitation particles continue to grow and eventually fall to the ground on the windward side of the mountain as precipitation. Water molecules leave the air parcel as precipitation. The precipitating water cannot further affect the parcel because the water molecules are now on the ground.

At the top of the mountain, the relative humidity of the parcel is 100%. As the air sinks, it warms, so the relative humidity decreases. The cloud particles evaporate as the air sinks and warms. The cloud disappears. With no cloud in the parcel, the parcel now warms all the way down the mountain at the dry adiabatic lapse rate. This is a crucial difference because on the way up the mountain the parcel cools for part of its journey at the (smaller) saturated adiabatic rate. As a result, the parcel ends up on the lee side warmer than it was at the very beginning.

In addition, there are fewer water molecules in the parcel because of the precipitation on the windward side. Increasing the temperature and removing water vapor from the air both act to lower the relative humidity. This is why regions located downwind of a mountain range are both warmer and drier than their windward counterparts. These are the rain-shadow regions. The rain-shadow effect is strongest when the wind is nearly perpendicular to the mountain range. In these cases, the windward side will have more cloud cover than the leeward side. The increased cloud cover will in turn reduce the annual temperature range (see Chapter 3) on the windward side compared to the leeward side. This is supported by weather observations on each side of the Cascade Mountains of Washington, which run north to south in a region of west-to-east winds (Table 4.4).

These observations demonstrate the influence of the topography on water in the atmosphere. They also illustrate the far-reaching consequences of the principles we have studied in the first four chapters of this text. Observations are central to the study of weather and climate—so important, in fact, that we devote the next chapter to how we sense the atmosphere.

PUTTING IT ALL TOGETHER

Summary

Water in the atmosphere exists as water vapor, clouds, and precipitation. We can measure water vapor in a variety of ways. Mixing ratio, vapor pressure, relative humidity, and dew point temperature are the most common "yardsticks" of water vapor. The vapor pressure of saturated (100% relative humidity) air increases rapidly as the air is warmed. Also, changing the amount of water vapor or the air temperature changes the relative humidity. In clear, calm, precipitation-free conditions, the relative humidity is usually highest at sunrise and lowest during the mid-afternoon.

A cloud is a suspension of water droplets, ice crystals, or both. Updrafts keep clouds aloft. The formation of a cloud requires water vapor, saturated air, and nuclei onto which the vapor can condense or deposit. These nuclei can be water droplets or aerosol

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particles. Clouds over land have more, and smaller, cloud droplets than clouds over the oceans.

Fog is a cloud at ground level. Fogs develop through two processes that lead to saturation. Air can cool to the dew point and become saturated, producing radiation fog, advection fog, or upslope fog. Air can also saturate via evaporation of water into the air. This produces steam fog or evaporation fog.

Most other clouds are formed when air cools to saturation as it is lifted. The four primary mechanisms for lifting air are orographic lifting, frontal lifting, convection, and convergence near the surface.

Moisture affects how a parcel's temperature changes as it rises. The saturated adiabatic lapse rate of about 6° C per kilometer is less than the dry adiabatic lapse rate because of latent heating resulting from condensation. This means that saturated air parcels cool less quickly than dry parcels as they rise. As a result, an environment with a lapse rate in-between the dry and saturated adiabatic lapse rates can be unstable but only for saturated air parcels. This is called a "conditionally unstable" environment and promotes the growth of clouds and thunderstorms.

Clouds can be classified as layered (*strato-*) or convective (*cumulo-*) and also as low, middle (*alto-*), high (*cirro-*), or precipitating (*nimbo-*). The 10 basic cloud types are cirrus, cirrostratus, cirrocumulus, altostratus, altocumulus, cumulus, stratus, stratocumulus, nimbostratus, and cumulonimbus (the thunderstorm cloud).

Clouds play a major role in the greenhouse effect. They warm the surface but also reflect sunlight, cooling the surface. Today, clouds tend to cool the planet, but this could change in the future.

Precipitation processes differ in warm and cold clouds. In warm water-only clouds, precipitation-sized particles grow by collision and coalescence of large and small water droplets. In cold clouds with ice crystals, accretion, aggregation, and the Bergeron-Wegener process cause precipitation. In the latter process, ice crystals attract water vapor more strongly than do liquid water drops in a temperature range commonly found in mid-latitude clouds.

The most common forms of precipitation are rain, snow, freezing rain, and sleet. Freezing rain and sleet form when there is a temperature inversion near the surface. The location of mountains and valleys affect the type and amount of both clouds and precipitation. In particular, regions downwind of a mountain range are drier and sunnier than the upwind slopes.

Key Terms

You should understand all of the following terms. Use the glossary and this chapter to improve your understanding of these terms.

Accretion
Advection fog
Aggregation
Altocumulus
Altostratus
Anvil
Bergeron-Wegener process
Cirrocumulus
Cirrostratus
Cirrus
Cloud droplet
Collision-coalescence
Column
Condensation nuclei
Conditionally unstable
Contact nucleation
Convection
Convective clouds
Convergence
Crystal habit
Cumulonimbus
Cumulus
Curvature effect
Dendrites
Deposition nucleation
Dew
Dew point
Dew point depression
Drizzle
Evaporation
Evaporation fog
Fallstreaks
Fog
Freezing nucleation
Freezing rain
Frontal fogs
Frontal lifting
Frost
Frost point
Graupel
Hail
Heterogeneous nucleation
Hexagonal plates
Homogeneous nucleation
Hydrophobic nuclei
Hygroscopic nuclei
Ice crystals
Ice nuclei
Immersion nucleation
Layered clouds
Lifting condensation level
Mammatus
Mixing ratio
Needle
Nimbostratus
Nucleation
Orographic lifting
Precipitation
Radiation fog
Rain
Rain shadow
Relative humidity
Saturated adiabatic lapse rate
Saturation
Saturation vapor pressure
Shower
Sleet
Snow
Snowflake
Solute effect
Steam fog
Stratocumulus
Stratus
Supercooled water
Supersaturation
Updraft
Upslope fog
Vapor pressure
Virga

Review Questions

1. Name two ways that you can cause a parcel of air to become saturated.
2. One day, the dew point is 20° C. The next day at the same location, the dew point is 10° C. On which day are there more water vapor molecules in the air? On which day is the relative humidity higher? (Hint: You may not have enough information to answer both questions.)
3. Why does the daily cycle of relative humidity look like the reverse of the daily cycle of temperature, with a maximum when the temperature is at a minimum? Hint: What's the definition of relative humidity, and how does one of the variables in the definition relate to temperature?
4. Do you think that relative humidity will reach its daily maximum at sunrise and be at its daily minimum at mid-afternoon on a day that is cloudy with rain in the afternoon?
5. On November 28, 2001, the University of Virginia played Michigan State University in a basketball game held at the Richmond Coliseum. It was an unusually warm, muggy evening. Underneath the basketball court was a sheet of ice used for hockey games. The game was halted because of slippery floor conditions. Using your understanding of water in the atmosphere, can you explain why the floor became slippery?
6. On a cold night when frost is predicted, you park your car underneath a tree instead of out in the open. Will frost form on your windshield? Explain why or why not.
7. Sometimes a fog will appear over a roadway after a summer rain shower. What type of fog is this?
8. A parcel of air at sea level has a temperature of 20° C and a dew point of 0° C. Assume that the dew point does not

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change as the unsaturated parcel rises. What will be the altitude of the lifting condensation level, where the parcel becomes saturated and the bottom of a puffy cloud forms?

9. Continuing from the previous question, what will the temperature of the air parcel be if it keeps rising to an altitude of 3 kilometers? What will the dew point be?
10. If an environment has a lapse rate of 8° C per kilometer, is it absolutely stable, absolutely unstable, or conditionally unstable? Could thunderstorms be a possibility in this situation? Is a temperature inversion present?
11. Why is there no such thing as a cirronimbus cloud?
12. Name the cloud type associated with each of the following: optical effects, thunderstorms, and fair weather.
13. Walking outside, you hold up your hand to the sky. You see a lumpy cloud that has features as big as your thumb. What is the name of this cloud?
14. Tor Bergeron observed that if a fog formed in a forest and the temperature was above 0° C (32° F), the fog extended down to the ground. If the temperature was below −5° C (23° F), the fog would not reach the forest floor. Explain his observation.
15. Explain how clouds help warm the ground. How do they help cool the ground? How does the altitude of the cloud affect its ability to warm or cool the ground?
16. An ice crystal grows for 5 minutes in a supersaturated environment with a temperature of −1° C. The crystal is carried to a different part of the cloud where the temperature is −14° C and the environment is still supersaturated. The crystal stays in this region of the cloud for another 5 minutes. Draw a picture of what the ice crystal might look like. Compare your picture to what you get using the Growing a Snowflake learning applet.
17. Many public restrooms have automatic hand-dryers. Why do they use hot air instead of cold air? The instructions say to place your hands in the airflow and gently rub them together. Explain how this dries your hands more rapidly than just holding them motionless in the air.
18. What temperature pattern must be present to cause freezing rain or sleet?
19. Can you easily dry your wet laundry outside when the temperature is below freezing? Explain your conclusion.
20. You are driving when very large raindrops suddenly splash onto your windshield. Is the updraft in the cloud above you strong or weak? Which cloud type is probably above you: nimbostratus or cumulonimbus?
21. What factors determine the growth of an ice crystal? How are these factors different than the factors that affect cloud droplet growth?
22. It starts snowing in very dry air that is above 0° C. Do you think the air temperature will rise, fall, or remain the same? (Hint: What will happen to the snow, and how will that cause an energy gain or loss by the atmosphere? Test your answer by using the Precipitation Types learning applet.)

Observation Activities

1. Throughout the course, either take photographs or keep a written log of the variety of cloud types you observe. Note the day, time, and general weather conditions at the time of your observation. Which cloud types were the hardest to identify or photograph, and why?
2. The purpose of this exercise is to relate the formation of bubbles in beer or soda (pop) to the formation of cloud droplets. Beer, which can certainly be non-alcoholic, tends to form better bubbles than soda. Pour the beer or soda into a clear glass. Where do the bubbles tend to form and why? What happens to the bubbles after they form? Pour some salt into the glass (sand or sugar can be used if salt isn't handy). What happens when you pour the salt in? Why does this happen?