Steven A. Ackerman and John A. Knox

THIRD EDITION

METEOROLOGY

Understanding the Atmosphere

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4 Water in the Atmosphere

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CHAPTER OUTLINE, CONTINUED

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- Ice Nuclei
- Cloud Particle Growth by Condensation and Deposition

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- Advection Fog
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LIFTING MECHANISMS THAT FORM CLOUDS

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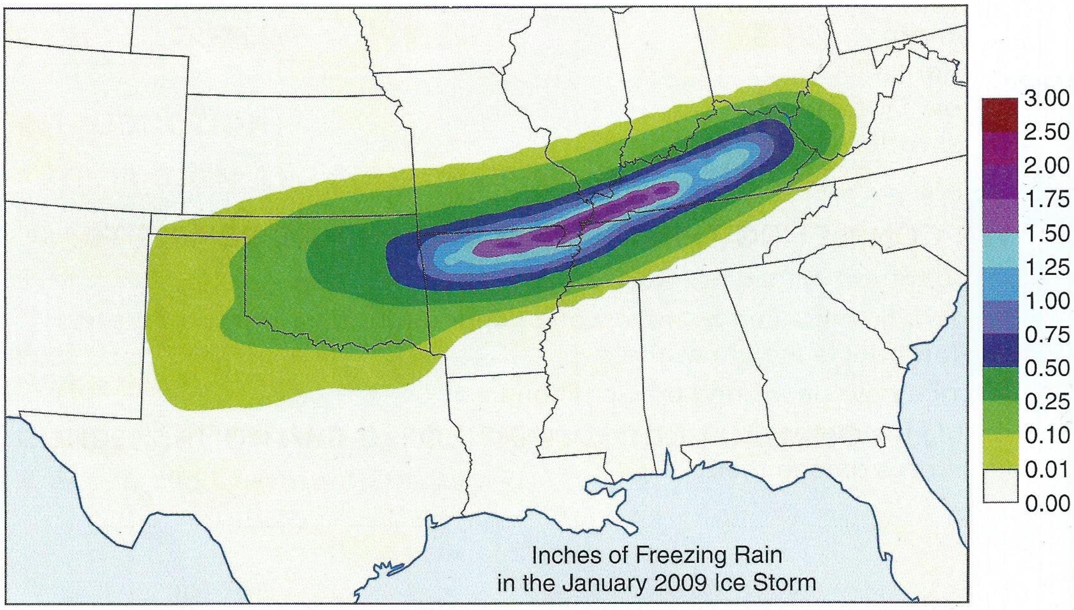
PUTTING IT ALL TOGETHER
- Summary
- Key Terms
- Review Questions
- Observation Activities

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

On January 26–28, 2009, a layer of ice encased the central United States from Texas to West Virginia. The precipitation fell mostly as rain, but the temperatures at the ground were cold enough for the rain to freeze on contact with anything it touched (see figures below). The storm coated many areas with more than an inch (2.5 cm) of ice, killing more than 60 people—35 in Kentucky alone. Most of the deaths were due to traffic accidents, extreme cold, and carbon monoxide poisoning (caused by power generators or kerosene heaters being used indoors without proper ventilation). Trees fell and power lines snapped under the weight of the ice, leaving more than 1.3 million people without electricity. It seemed as if Mother Nature had declared an icy war on the region—and as if to confirm this impression, the entire Kentucky Army National Guard was mobilized to help with the many problems left in the wake of this storm.

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



Source: NOAA/NWS.

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 understand evaporation better, 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 4-1a**). 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 4-1b).

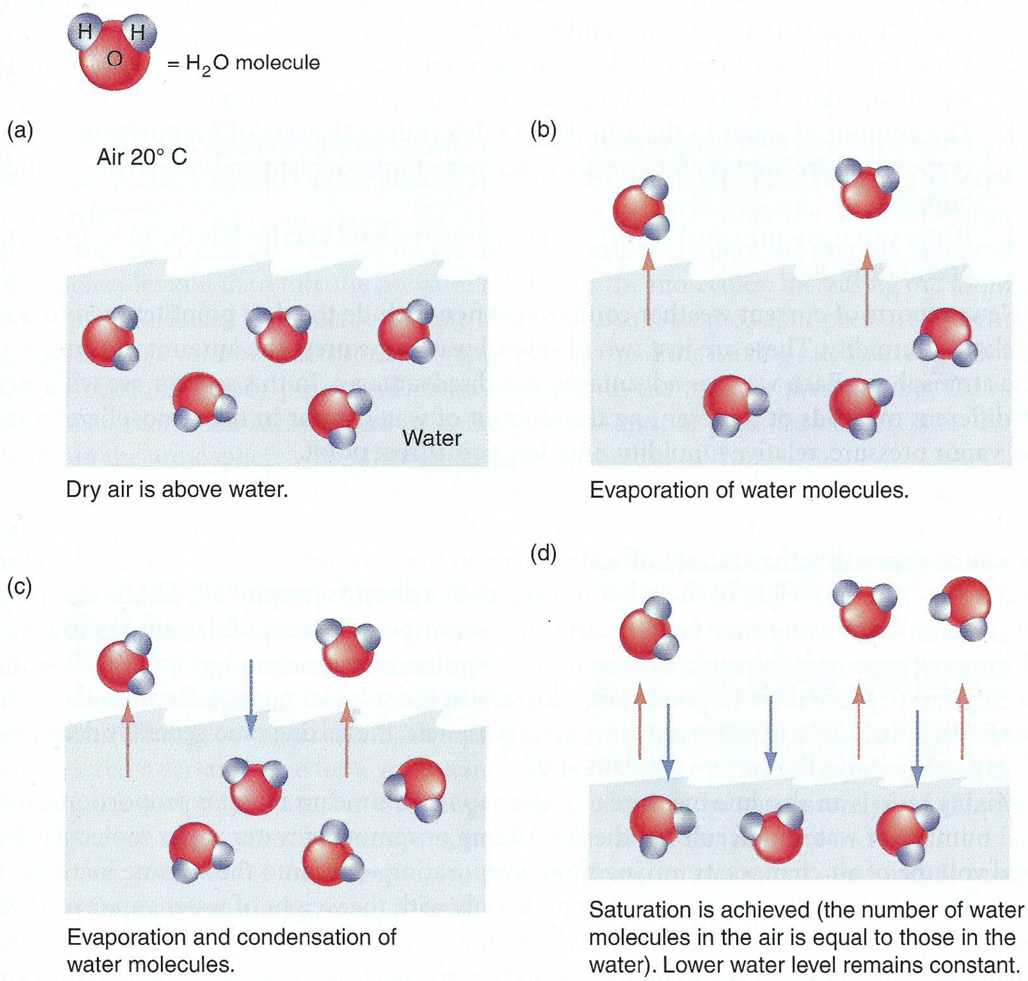


FIGURE 4-1 The sequence of events that leads to saturation of air. (a) Initially, dry air lies above water at 20° C. (b) Evaporation into the air begins. (c) Condensation back to the water occurs; however, evaporation exceeds condensation and the number of water molecules in the air increases. The air is saturated in (d), where the number of evaporating and condensing water molecules is equal.

As time goes on, some water molecules at the liquid surface will be escaping and evaporating. Other molecules in the air will be captured (Figure 4-1c). 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 4-1d).

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 do not require counting molecules. We look at these different methods in the following sections.

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 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. Because 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.

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

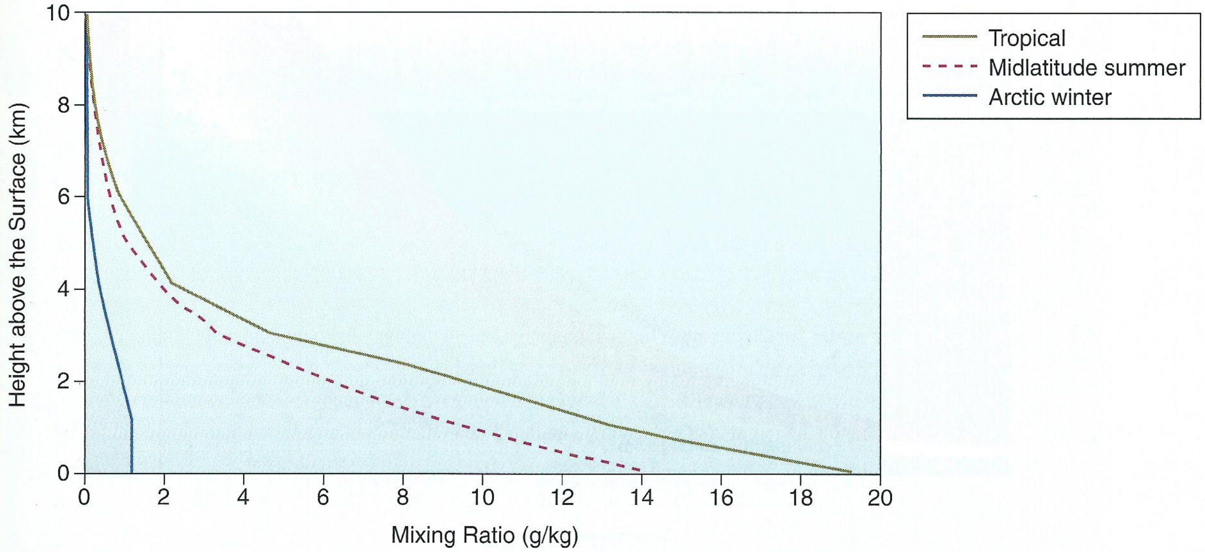


FIGURE 4-2 This graph shows the vertical distribution of the mixing ratio for three different atmospheric conditions. Atmospheric water vapor comes from the surface of Earth. This explains why values of mixing ratio usually decrease with altitude even at different regions of the world and during different seasons.

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. Increasing the air temperature will increase the vapor pressure. Changing the air temperature changes the average kinetic energy of the molecules and therefore the pressure exerted by the molecules. 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 often 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 4-1d), 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 ridge top 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.

"Moisture Graph" to learn how moisture affects static stability.

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, in

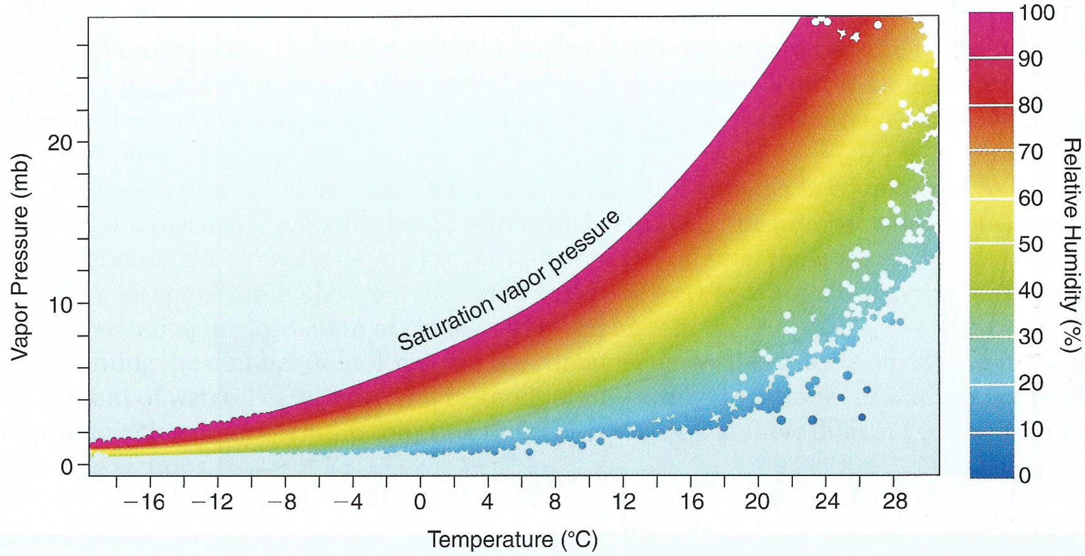


FIGURE 4-3 Observations of vapor pressure as a function of temperature on a ridge top at Black Rock Forest along the lower Hudson River in New York. The observations were made hourly from December 1994 through mid April 2001. Relative humidity is indicated by the color coding. Notice how the highest 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. (*Source*: Black Rock Forest Consortium.)

equilibrium between the air and the surface beneath it. 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.**

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 that 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 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 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 degrees 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.

Heat Index Values (°F)

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

- 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.

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

(continued)

Box 4-1 Atmospheric Moisture and Your Health, continued

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 discussed 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 and is much more uncontrollable than in rain. Even off the roadways, you can easily fall and break bones on slippery steps or a sidewalk. Overall, in the United States each year, approximately 7000 highway deaths and 450,000 injuries are associated with poor-weather–related driving conditions, according to Congressional testimony by Dr. Richard Anthes, President of the University Corporation for Atmospheric Research in Boulder, CO. This means that weather plays a role in about 28% of all crashes and 19% of all highway fatalities. 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.

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.

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.

"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.

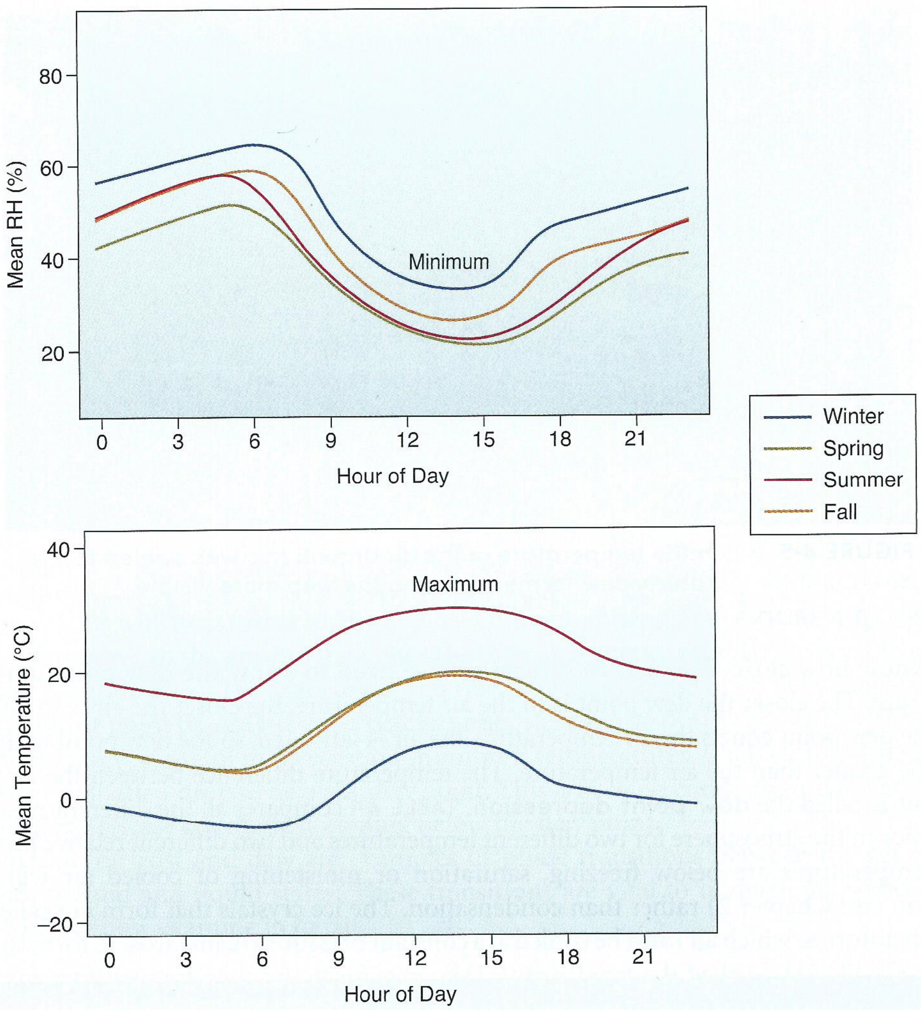


FIGURE 4-4 A climatology of hourly temperature and relative humidity, as observed at Bandelier National Monument, New Mexico, from October 1988 through May 1999. In the usually dry, clear climate of New Mexico, the daily cycle of relative humidity is exactly the opposite of the temperature cycle, with 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. (Data from Diurnal Cycle. Retrieved December 10, 2010, from http://vista.cira.colostate.edu/improve/data/graphicviewer/diurnal.htm.)

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 a glass of an ice-cold drink. 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.

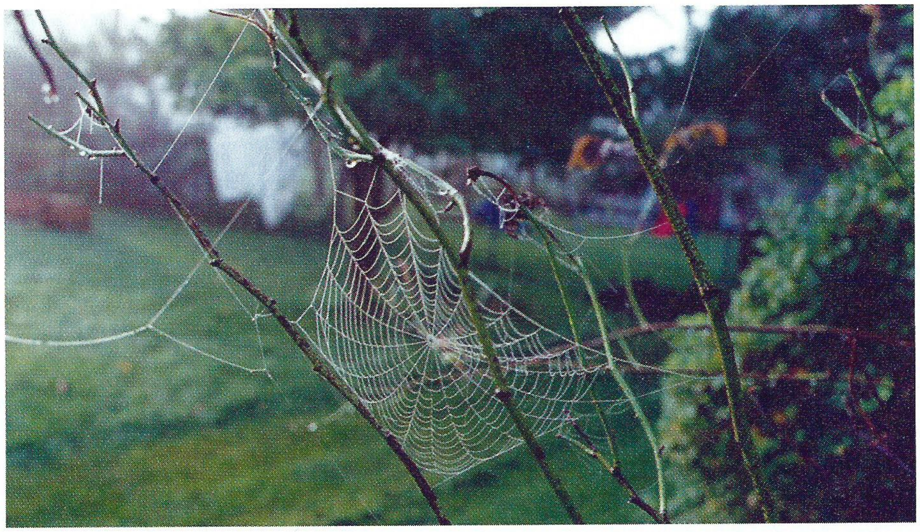


FIGURE 4-5 When the temperature of the air around this web cooled to the dew point temperature, dew formed, making the web more visible.

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 (see Chapter 2) rather than condensation. The ice crystals that form are called **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. Yet another frozen water type is **rime**, which is a white, opaque deposit of ice formed by the rapid freezing of water drops as they collide with an object at or below freezing.

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.

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

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.

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.

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

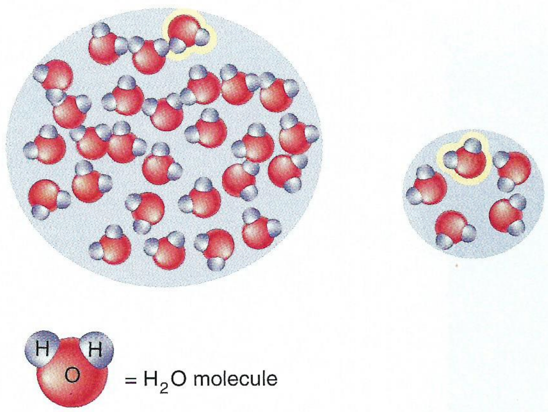


FIGURE 4-7 The smaller the drop, the more curved the surface, reducing the number of neighbors for each water molecule at the surface. This curvature effect makes it easier for small drops to evaporate.

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 are not 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.

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 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. Brewers sterilize and clean the bottles to keep the beverage from going bad. 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 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 molecules 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 large body of water, can be supercooled.

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 the volume of water, the greater the chances that a few of the molecules will align in the proper manner to form ice when the temperature falls below freezing. 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. After 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

After 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 and the air is saturated, 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**.

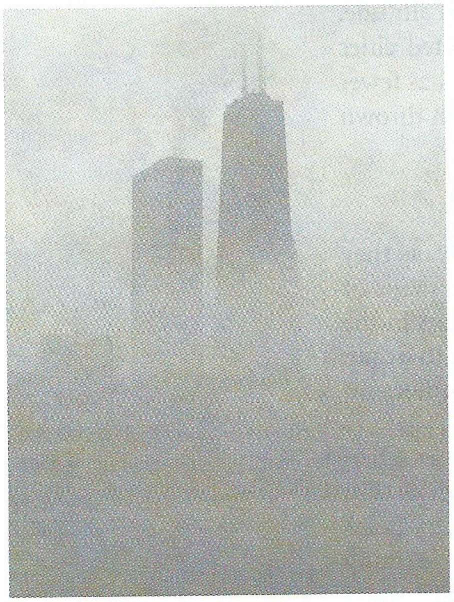


FIGURE 4-8 Fog consists of tiny water droplets that can reduce visibility to less than 1 kilometer (0.6 miles).

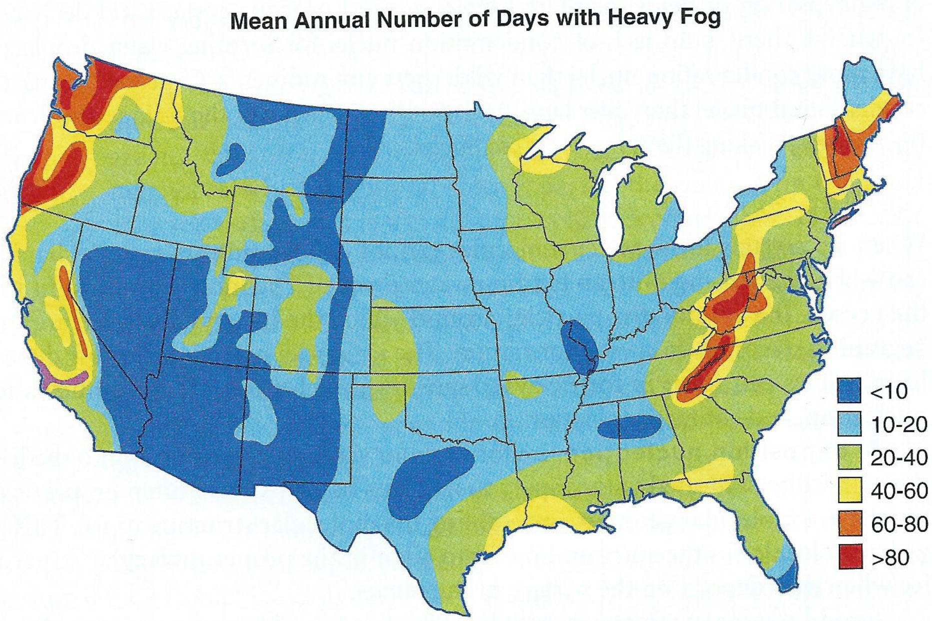


FIGURE 4-9 The mean annual number of dense fog days with visibility less than 300 meters across the United States.

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.

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, keeping 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).

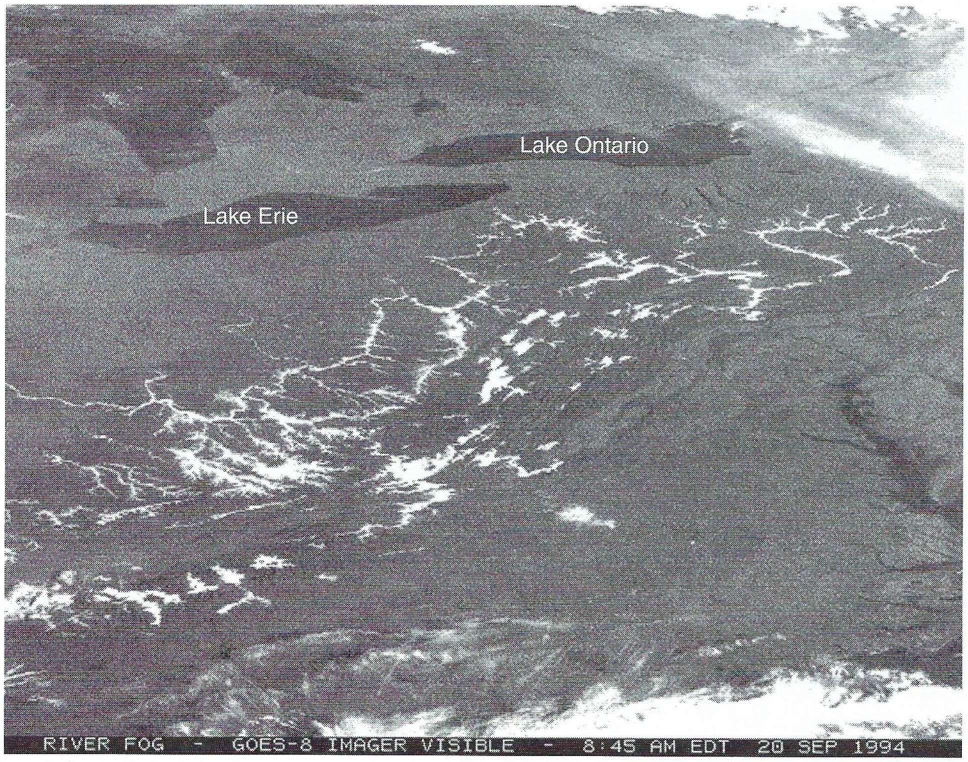


FIGURE 4-10 Satellite photograph on the morning of September 20, 1994, showing radiation fog (narrow white areas) in the Ohio River valley and its tributaries.

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, forming a fog. 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.

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.

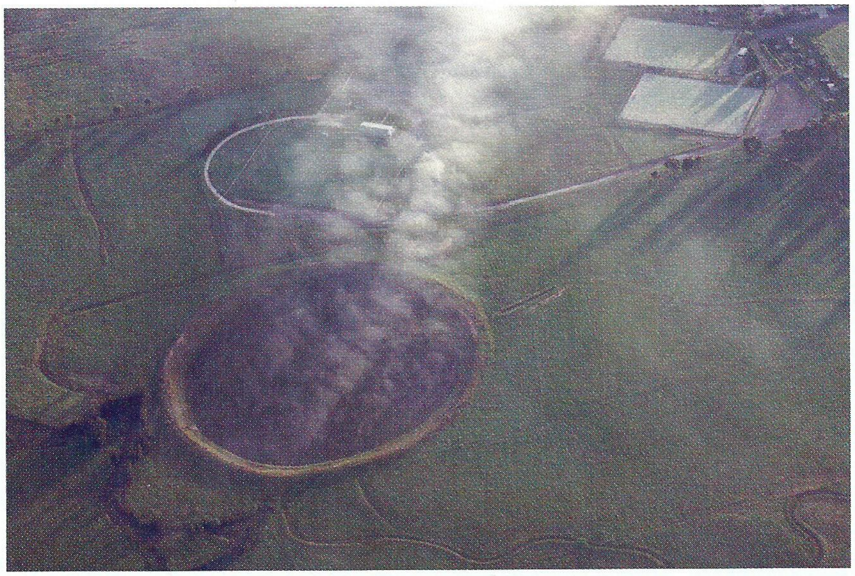


FIGURE 4-12 An aerial view of steam fog rising and moving downwind from a small lake in Australia. Using our knowledge of relative humidity plus the shadows in this photograph, can you determine what time of day this picture was taken?

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 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 form only 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 midlatitude regions of the globe. Steam fog is common when very cold air rushes over unfrozen 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.

LIFTING MECHANISMS THAT FORM CLOUDS

"Cloud Base Altitude" to explore the relationship between temperature, dew point, and the base of a cumulus cloud.

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.

FIGURE 4-13 depicts four mechanisms that cause air to ascend. Air is lifted as it moves against a mountain range (Figure 4-13a). The air cannot go through the mountain, and so it flows over the mountain. This is **orographic lifting.**

Other lifting mechanisms can also cause clouds to form. For example, at the same pressure, cold air is denser 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 4-13b). 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 4-13c).

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 4-13d). The opposite of convergence is **divergence**, which is the horizontal spreading out of air.

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 section, we examine the role of moisture and stability in cloud development.

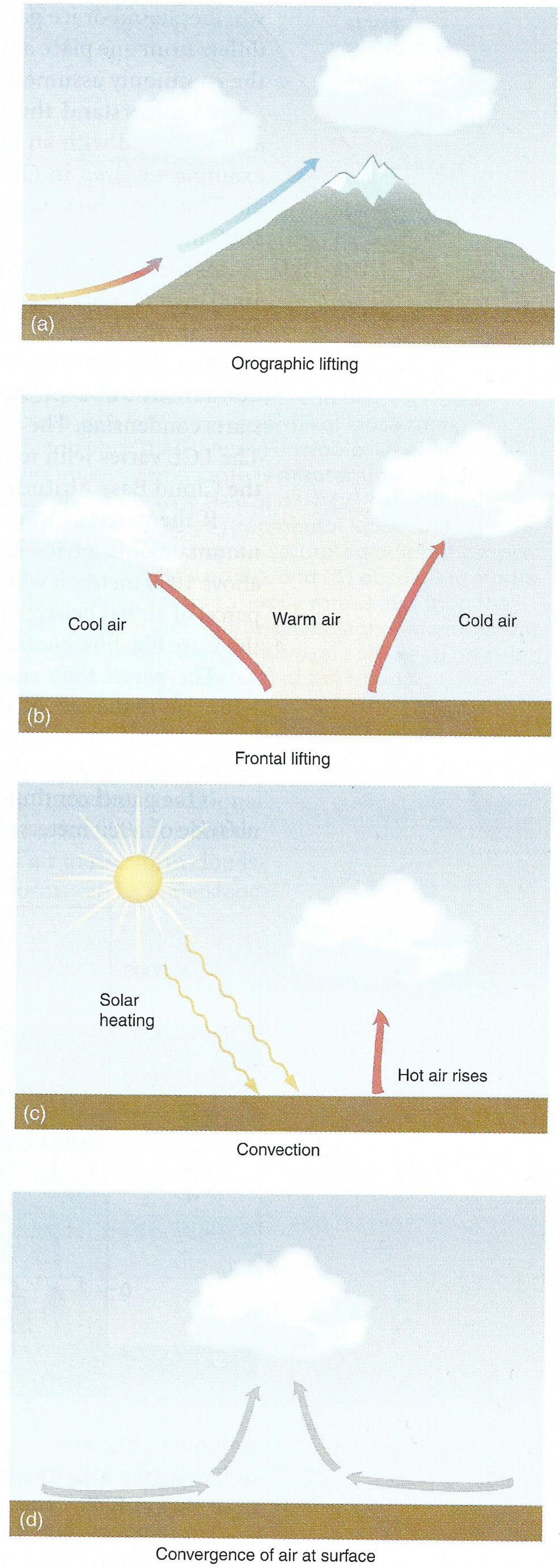


FIGURE 4-13 Depiction of the four mechanisms that cause air to ascend and form a cloud.

STATIC STABILITY AND CLOUD DEVELOPMENT

We introduced the concept of static stability in Chapter 3. 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 with 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., it is not unsaturated). Therefore, we have to modify our understanding of static stability to take phase changes of 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 the commonly assumed saturated adiabatic lapse rate of 6° C per kilometer.

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.

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

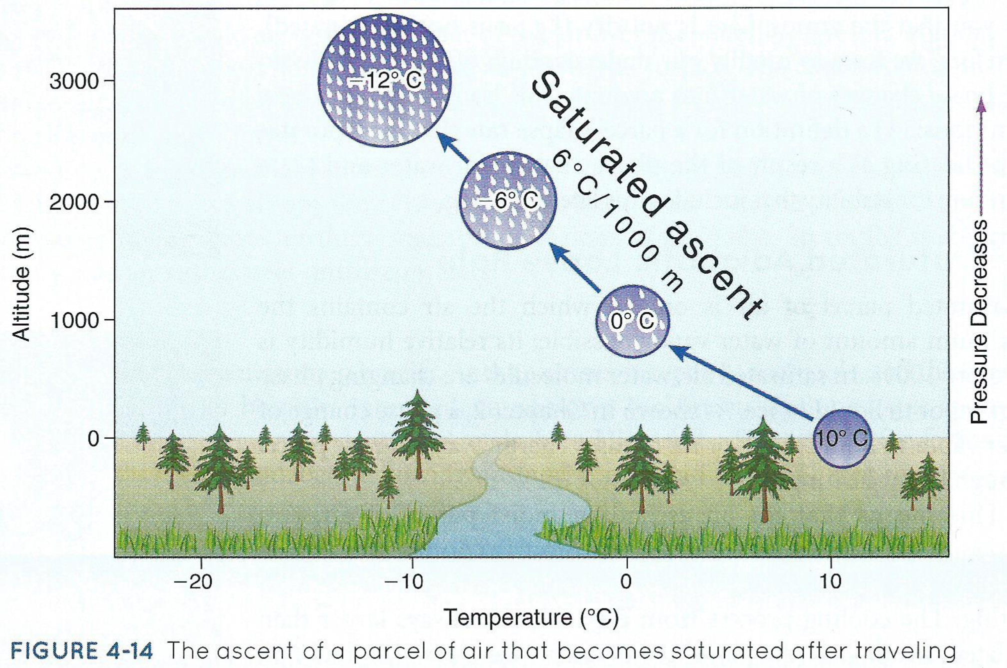


FIGURE 4-14 The ascent of a parcel of air that becomes saturated after traveling upward 1 km. The rate of cooling during ascent is less than the dry adiabatic lapse rate because of the warming as a result of latent heating, as water vapor molecules condense. A typical saturated adiabatic lapse rate is 6° C per kilometer, but the exact value varies. Compare this figure with Figure 3-19, which depicts the ascent of an unsaturated air parcel. How do the two differ?

−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 same thing as saying that the saturated adiabatic 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: 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 rising saturated air parcel will become progressively warmer than its environment as it rises and therefore will keep rising. In such unstable situations, tall clouds can form, especially thunderstorms.

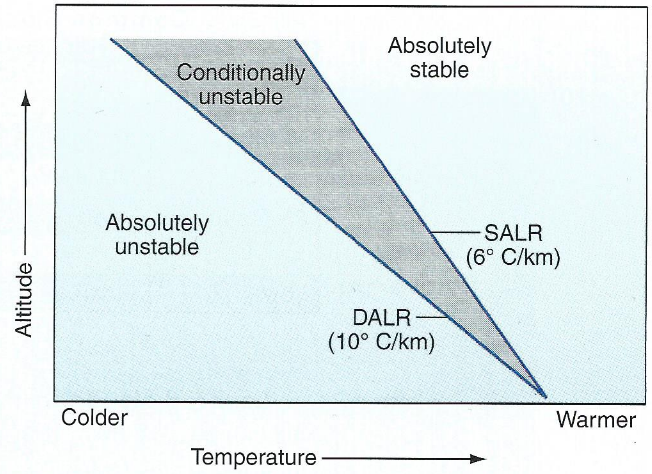


FIGURE 4-15 An environment can be described in three different ways in terms of lapse rates and stability: (1) Absolutely unstable, when the environmental lapse rate is greater than the dry adiabatic lapse rate (DALR; left); (2) conditionally unstable, when the environmental lapse rate is in between 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 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	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 convective—and their altitude. Clouds are also categorized based on whether or not they are precipitating.

TABLE 4-3 Common Cloud Types

Cloud Type	Layered Cloud	Convective Cloud	Mixed/ Neither	Typical Altitudes 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

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 that is forced to rise.

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 are 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.

"Name That Cloud" to quiz yourself on the cloud types discussed in this section.

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.

Low Clouds

Stratus

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 is sometimes visible on car windshields during thick stratus.

Stratocumulus

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

Precipitation typically 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.

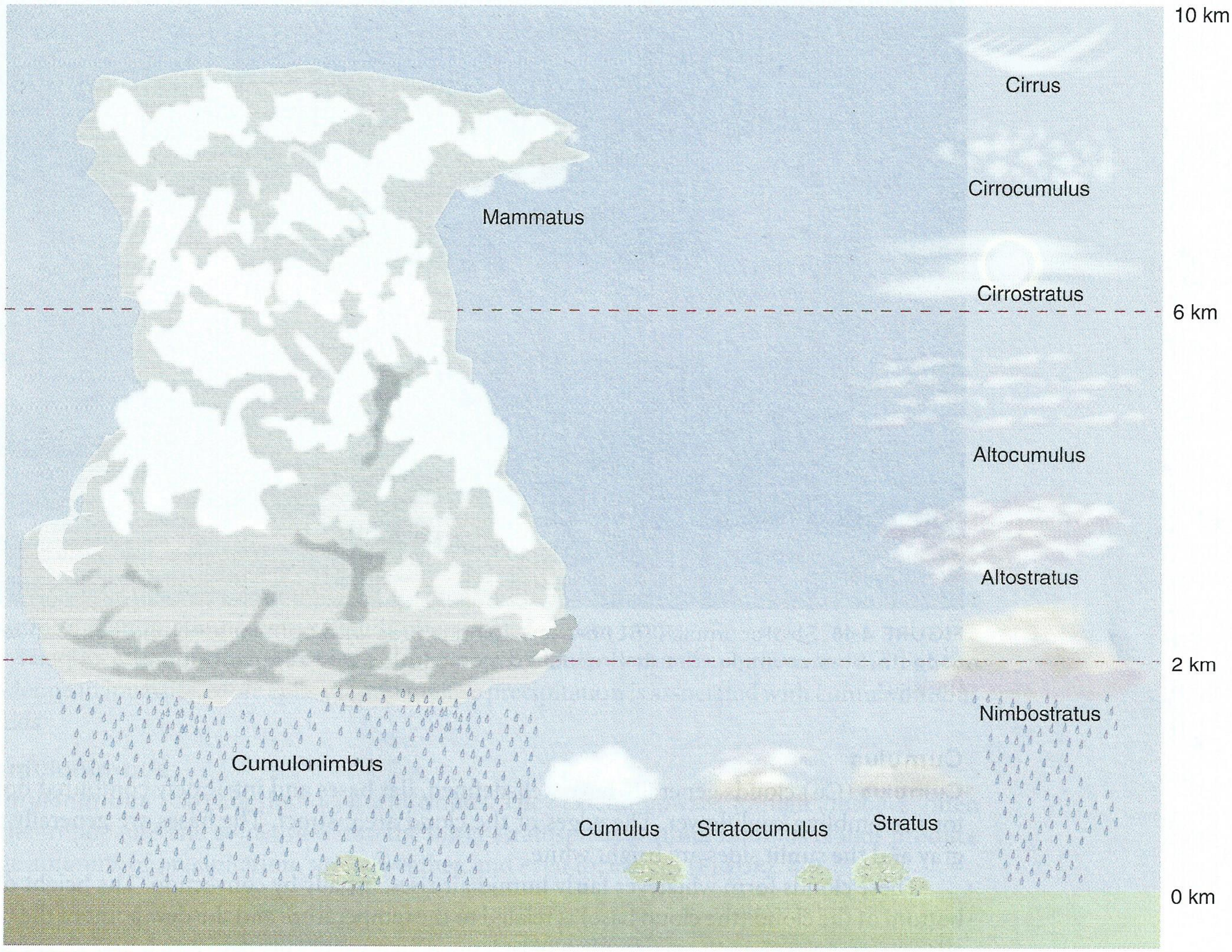


FIGURE 4-16 The major cloud types arranged by their typical altitude.

FIGURE 4-17 Stratus (St) are low-altitude layered clouds.

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.

Cumulus

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 southwestern 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 such as 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.

Precipitating Clouds

Nimbostratus

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

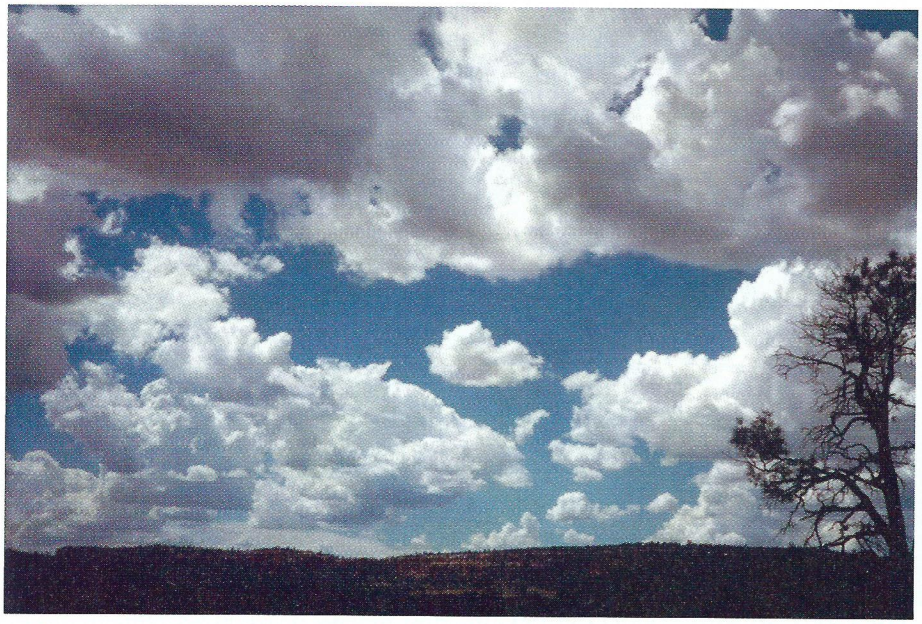


FIGURE 4-19 Cumulus (Cu) clouds are often observed on summer days.

the cloud. For this reason, nimbostratus sometimes look similar to stratus, stratocumulus, or altostratus clouds. Nimbostratus clouds often precede warm fronts.

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

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).

FIGURE 4-20 Nimbostratus (Ns) are deep layered clouds that bring precipitation and appear dark gray to pale blue in color.

FIGURE 4-21 Cumulonimbus (Cb) over Lake Wingra, Madison, Wisconsin.

A distinguishing feature of cumulonimbus is the flattened **anvil** shape of the top of the cloud. The anvil develops when the updraft slows 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, and see this book's cover). 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

FIGURE 4-22 Pouchy mammatus clouds (top half of photograph) sometimes form on the underside of cumulonimbus anvils.

develop into an organized system, the chance of severe weather often increases, as we will see in Chapter 11.

Middle Clouds

Altostratus

Altostratus (As) are layered clouds made up mostly of liquid water droplets. They are gray to pale blue in appearance (FIGURE 4-23). Altostratus form when the middle layers of the 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 often observed ahead of a warm front, before the nimbostratus.

Altocumulus

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). Altocumulus clouds often appear ahead of a warm front, prior to altostratus. If other cloud types accompany altocumulus, a storm is probably approaching.

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

High Clouds

Cirrocumulus

Cirrocumulus (Cc) clouds are thin, white clouds that often appear in ripples arranged in a regular formation (FIGURE 4-25). The smaller size of the individual cumulus lumps in cirrocumulus

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.

FIGURE 4-24 Altocumulus (Ac) occur in the middle levels of the atmosphere when the air is moist.

FIGURE 4-25 Cirrocumulus (Cc) are thin, white clouds that appear high in the troposphere.

clouds distinguish this cloud type from altocumulus. Cirrocumulus clouds are 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 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.

Cirrostratus

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 4-26**). 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

Cirrus (Ci) are wispy, fibrous, white clouds that are made of ice crystals. They often occur as wisps here and there across the sky and are aligned in the same direction as the upper-level winds (**FIGURE 4-27**). 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

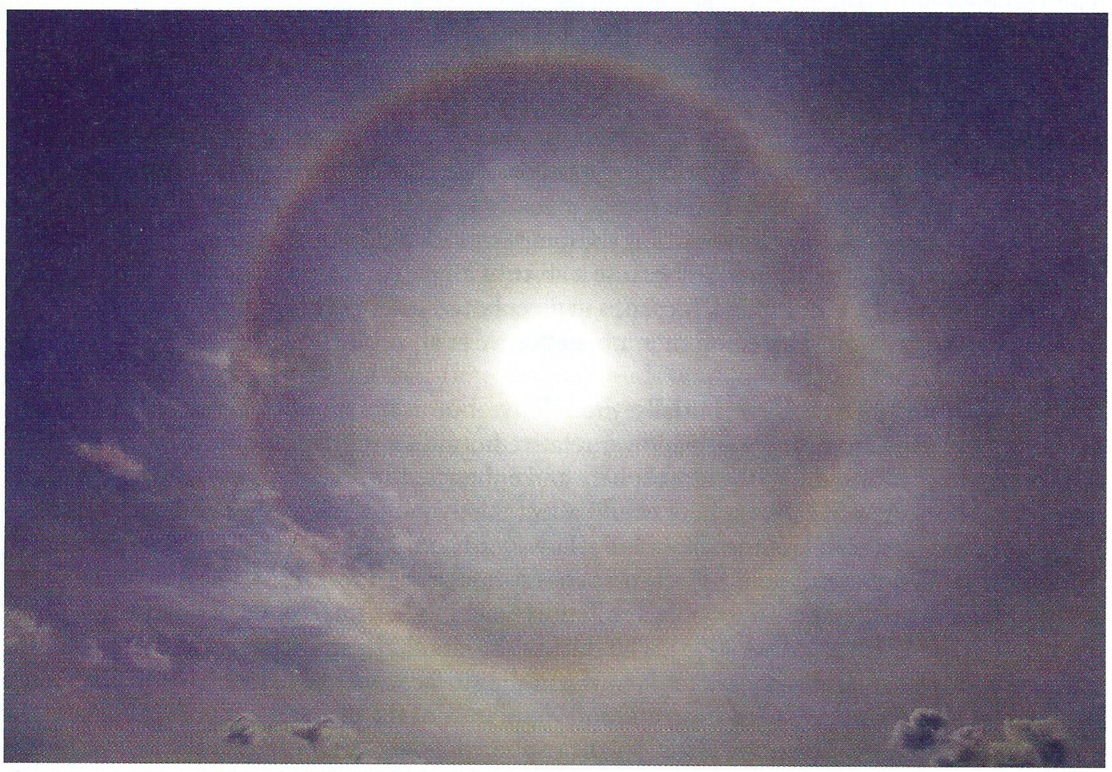


FIGURE 4-26 Cirrostratus (Cs) are layered clouds that are sometimes observed in connection with optical effects, such as halos and sundogs.

FIGURE 4-27 Cirrus (Ci) are wispy, fibrous, white clouds that are composed of ice crystals.

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.

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 result in a climate change because the atmosphere becomes more effective at absorbing longwave energy emitted by the surface.

As 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 4-28**). 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

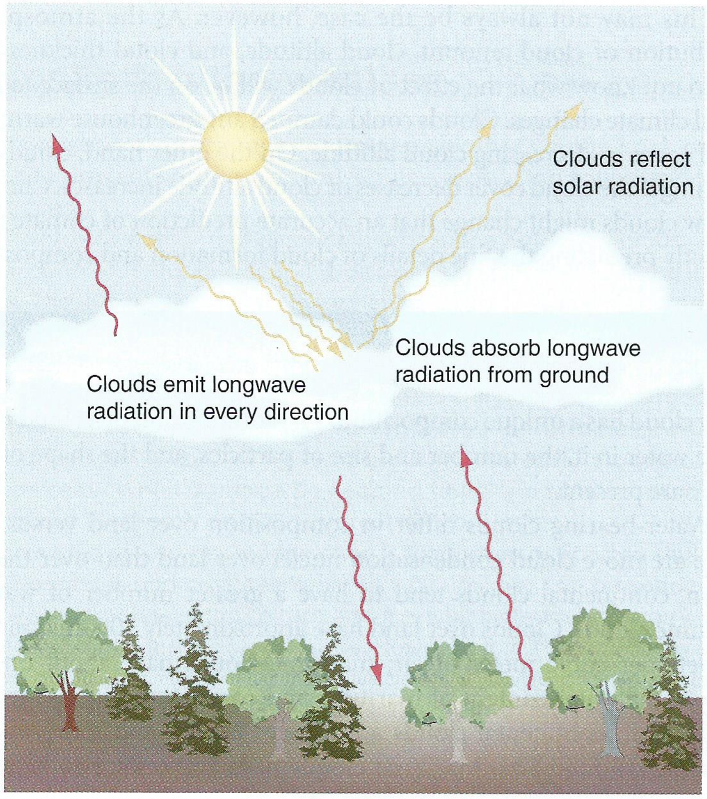


FIGURE 4-28 In the solar spectrum, clouds tend to cool Earth. In the longwave spectrum, they tend to warm the planet.

form of longwave radiation (Figure 4-28). 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 out to space 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, which keep that energy from being lost to space. Thus, with respect to longwave radiation losses to space, cirrus clouds tend to warm the planet. The longwave effect dominates, and cirrus clouds, in general, tend to warm the planet in comparison to clear-sky conditions.

Stratus clouds over water tend to cool the planet. This is because stratus clouds are very effective at reflecting solar energy out to space reducing the net energy gain. Stratus are low in the atmosphere and therefore have temperatures that are close to the surface temperatures, so adding them to clear sky conditions does not change the outgoing longwave energy. The shortwave effect dominates, and maritime stratus clouds tend to cool the planet.

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.

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.

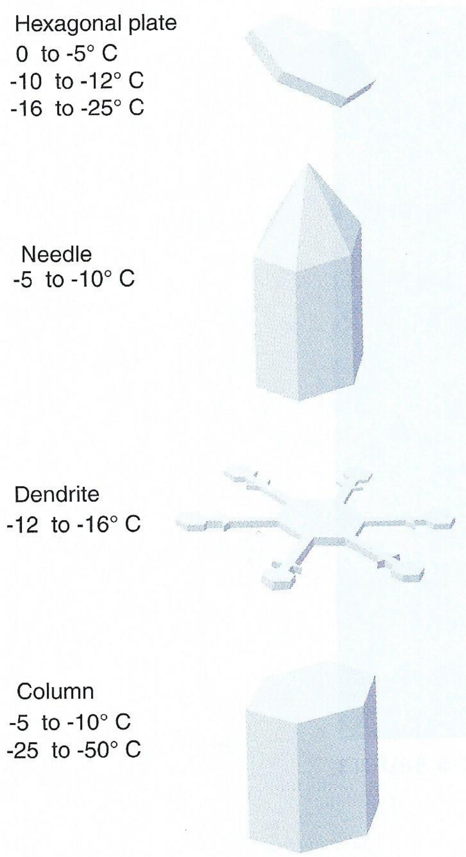


FIGURE 4-29 The four basic ice crystal habits are column, needle, hexagonal plate, and dendrite. The shape in which an ice particle grows depends on the temperature of its environment. Try this out for yourself by using the "Growing a Snowflake" learning applet.

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 as 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 4-29 shows the four basic shapes of ice crystals, each of which occur preferentially in the following temperature ranges: the **hexagonal plate** (0° C to –5° C; –10° C to –12° C; –16° C to –25° C), the **needle** (–5° C to –10° C), the **column** (–5° C to –10° C; –25° C to –50° C), and the **dendrite** (–12° C 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.

"Growing a Snowflake" make your own beautiful and complex ice crystals.

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, returning it to the Earth's surface. Rain, snow, sleet, freezing rain, and hail are all forms of precipitation.

"Precipitation Formation" explore how cloud particles become precipitation size.

Dew and frost also remove water vapor through condensation or deposition onto surfaces on the ground. Dew and frost are not precipitation 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. These growth processes are dependent on the cloud temperature. **Warm clouds** are those that have temperature greater than freezing throughout the cloud. **Cold clouds** have temperatures that are below freezing. After discussing how particles grow into precipitation, we examine 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.)