

**Meteorology Today: An Introduction to
Weather, Climate, and the Environment**

Ninth Edition

C. Donald Ahrens

Development Editor: Jake Warde

Assistant Editor: Liana Monari

Editorial Assistant: Paige Leeds

Technology Project Manager: Alexandria Brady

Marketing Manager: Joe Rogove

Marketing Assistant: Elizabeth Eong

Marketing Communications Manager:
Belinda Krohmer

Project Manager, Editorial Production:
Hal Humphrey

Art Director: Vernon Boes

Print Buyer: Rebecca Cross

Permissions Editor:
Margaret Chamberlain-Gaston

Production Service: Janet Bollow Associates

Text Designer: Janet Bollow

Art Editor: Janet Bollow

Copy Editor: Stuart Kenter

Illustrator: Charles Preppernau

Cover Designer: William Stanton

Cover Image: Copyright R. Hoelzl/
Peter Arnold, Inc.

Compositor: Graphic World, Inc.

© 2009, 2007 Brooks/Cole, Cengage Learning

ALL RIGHTS RESERVED. No part of this work covered by the copyright herein may be reproduced, transmitted, stored or used in any form or by any means graphic, electronic, or mechanical, including but not limited to photocopying, recording, scanning, digitizing, taping, Web distribution, information networks, or information storage and retrieval systems, except as permitted under Section 107 or 108 of the 1976 United States Copyright Act, without the prior written permission of the publisher.

For product information and technology assistance, contact us at
Cengage Learning Customer & Sales Support, 1-800-354-9706

For permission to use material from this text or product,
submit all requests online at **cengage.com/permissions**

Further permissions questions can be emailed to
permissionrequest@cengage.com

Library of Congress Control Number: 2008928602

ISBN-13: 978-0-495-55573-5

ISBN-10: 0-495-55573-8

Brooks/Cole

10 Davis Drive
Belmont, CA 94002
USA

Cengage Learning is a leading provider of customized learning solutions with office locations around the globe, including Singapore, the United Kingdom, Australia, Mexico, Brazil, and Japan. Locate your local office at:

international.cengage.com/region

Cengage Learning products are represented in Canada by Nelson Education, Ltd.

For your course and learning solutions, visit **academic.cengage.com**

Purchase any of our products at your local college store or at our preferred
online store **www.ichapters.com**

Over land, a considerable amount of water vapor is added to the atmosphere through evaporation from the soil, lakes, and streams. Even plants give up moisture by a process called *transpiration*. The water absorbed by a plant's root system moves upward through the stem and emerges from the plant through numerous small openings on the underside of the leaf. In all, evaporation and transpiration from continental areas amount to only about 15 percent of the nearly 1.5 billion billion gallons of water vapor that annually evaporate into the atmosphere; the remaining 85 percent evaporates from the oceans. If all of this water vapor were to suddenly condense and fall as rain, it would be enough to cover the entire globe with 2.5 centimeters, (or 1 inch) of water.* The total mass of water vapor stored in the atmosphere at any moment adds up to only a little over a week's supply of the world's precipitation. Since this amount varies only slightly from day to day, the hydrologic cycle is exceedingly efficient in circulating water in the atmosphere.

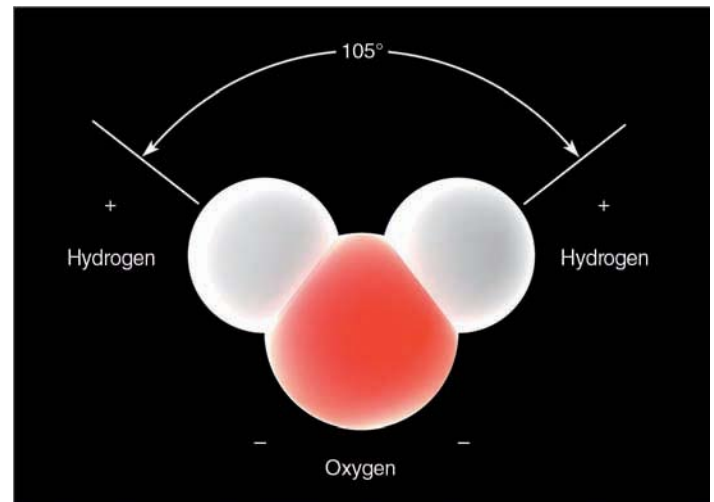
The Many Phases of Water

If we could see individual water molecules, we would find that, in the lower atmosphere, water is everywhere. If we could observe just one single water molecule by magnifying it billions of times, we would see an H_2O molecule in the shape of a tiny head that somewhat resembles Mickey Mouse (see • Fig. 4.2). The bulk of the “head” of the molecule is the oxygen atom. The “mouth” is a region of excess negative charge. The “ears” are partially exposed protons of the hydrogen atom, which are regions of excess positive charge.

When we look at many H_2O molecules, we see that, as a gas, water vapor molecules move about quite freely, mixing well with neighboring atoms and molecules (see • Fig. 4.3). As we learned in Chapter 2, the higher the temperature of the gas, the faster the molecules move. In the liquid state, the water molecules are closer together, constantly jostling and bumping into one another. If we lower the temperature of the liquid, water molecules would move slower and slower until, when cold enough, they arrange themselves into an orderly pattern with each molecule more or less locked into a rigid position, able to vibrate but not able to move about freely. In this solid state called *ice*, the shape and charge of the water molecule helps arrange the molecules into six-sided (hexagonal) crystals.

As we observe the ice crystal in freezing air, we see an occasional molecule gain enough energy to break away from its neighbors and enter into the air above. The molecule changes from an ice molecule directly into a vapor molecule without passing through the liquid state. This ice-to-vapor phase change is called **sublimation**. If a water vapor molecule should attach itself to the ice crystal, the vapor-to-ice phase change is called **deposition**. If we apply warmth to the ice

*If the water vapor in a column of air condenses and falls to the earth as rain, the depth of the rain on the surface is called *precipitable water*.



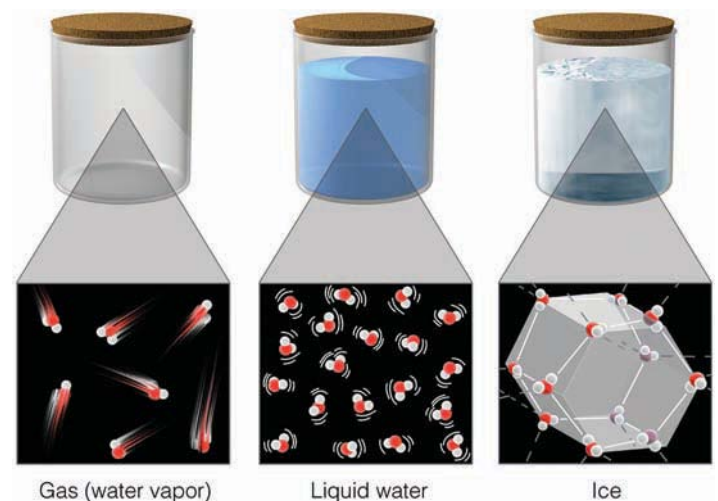
• FIGURE 4.2 The water molecule.

crystal, its molecules would vibrate faster. In fact, some of the molecules would actually vibrate out of their rigid crystal pattern into a disorderly condition—that is, the ice melts.

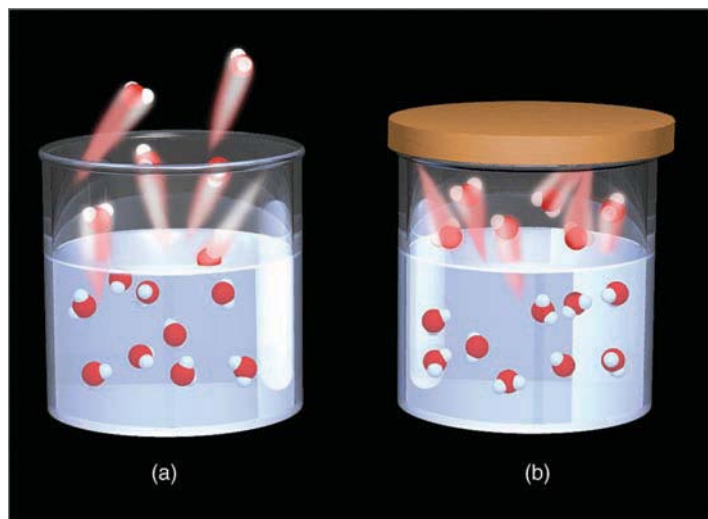
And so water vapor is a gas that becomes visible to us only when millions of molecules join together to form tiny cloud droplets or ice crystals. In this process—known as a *change of state* or, simply, *phase change*—water only changes its disguise, not its identity.

Evaporation, Condensation, and Saturation

Suppose we were able to observe individual water molecules in a beaker, as illustrated in • Fig. 4.4a. What we would see are water molecules jiggling, bouncing, and moving about. However, we would also see that the molecules are not all moving



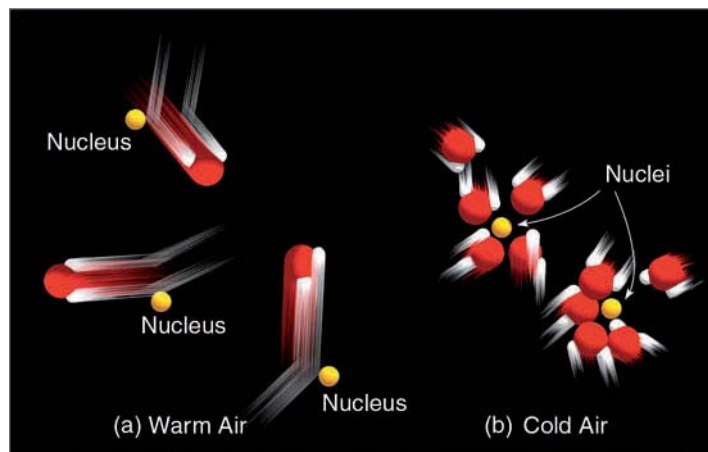
• FIGURE 4.3 The three states of matter. Water as a gas, as a liquid, and as a solid.



● **FIGURE 4.4** (a) Water molecules at the surface of the water are evaporating (changing from liquid into vapor) and condensing (changing from vapor into liquid). Since more molecules are evaporating than condensing, net evaporation is occurring. (b) When the number of water molecules escaping from the liquid (evaporating) balances those returning (condensing), the air above the liquid is saturated with water vapor. (For clarity, only water molecules are illustrated.)

at the same speed—some are moving much faster than others. At the surface, molecules with enough speed (and traveling in the right direction) would occasionally break away from the liquid surface and enter into the air above. These molecules, changing from the *liquid state into the vapor state* are *evaporating*. While some water molecules are leaving the liquid, others are returning. Those returning are *condensing* as they are changing from a *vapor state to a liquid state*.

When a cover is placed over the beaker (see Fig. 4.4b), after a while the total number of molecules escaping from the liquid (evaporating) would be balanced by the number re-



● **FIGURE 4.5** Condensation is more likely to occur as the air cools. (a) In the warm air, fast-moving H_2O vapor molecules tend to bounce away after colliding with nuclei. (b) In the cool air, slow-moving vapor molecules are more likely to join together on nuclei. The condensing of many billions of water molecules produces tiny liquid water droplets.

turning (condensing). When this condition exists, the air is said to be **saturated** with water vapor. For every molecule that evaporates, one must condense, and no net loss of liquid or vapor molecules results.

If we remove the cover and blow across the top of the water, some of the vapor molecules already in the air above would be blown away, creating a difference between the actual number of vapor molecules and the total number required for *saturation*. This would help prevent saturation from occurring and would allow for a greater amount of evaporation. Wind, therefore, enhances evaporation.

The temperature of the water also influences evaporation. All else being equal, warm water will evaporate more readily than cool water. The reason for this phenomenon is that, when heated, the water molecules will speed up. At higher temperatures, a greater fraction of the molecules have sufficient speed to break through the surface tension of the water and zip off into the air above. Consequently, the warmer the water, the greater the rate of evaporation.

If we could examine the air above the water in Fig. 4.4b, we would observe the water vapor molecules freely darting about and bumping into each other as well as neighboring molecules of oxygen and nitrogen. When these gas molecules collide, they tend to bounce off one another, constantly changing in speed and direction. However, the speed lost by one molecule is gained by another, and so the average speed of all the molecules does not change. Consequently, the temperature of the air does not change. Mixed in with all of the air molecules are microscopic bits of dust, smoke, salt, and other particles called *condensation nuclei* (so-called because water vapor condenses on them). In the warm air above the water, fast-moving vapor molecules strike the nuclei with such impact that they simply bounce away (see ● Figure 4.5a). However, if the air is chilled (Fig. 4.5b), the molecules move more slowly and are more apt to stick and condense to the nuclei. When many billions of these vapor molecules condense onto the nuclei, tiny liquid cloud droplets form.

We can see then that condensation is more likely to happen as the air cools and the speed of the vapor molecules decreases. As the air temperature increases, condensation is less likely because most of the molecules have sufficient speed (sufficient energy) to remain as a vapor. As we will see in this and other chapters, *condensation occurs primarily when the air is cooled*.*

Even though condensation is more likely to occur when the air cools, it is important to note that no matter how cold the air becomes, there will always be a few molecules with sufficient speed (sufficient energy) to remain as a vapor. It should be apparent, then, that with the same number of water vapor molecules in the air, saturation is more likely to occur in cool air than in warm air. This idea often leads to the statement that “warm air can hold more water vapor molecules before becoming saturated than can cold air” or, simply,

*As we will see later, another way of explaining why cooling produces condensation is that the saturation vapor pressure decreases with lower temperatures.

“warm air has a greater capacity for water vapor than does cold air.” At this point, it is important to realize that although these statements are correct, the use of such words as “hold” and “capacity” are misleading when describing water vapor content, as air does not really “hold” water vapor in the sense of making “room” for it.

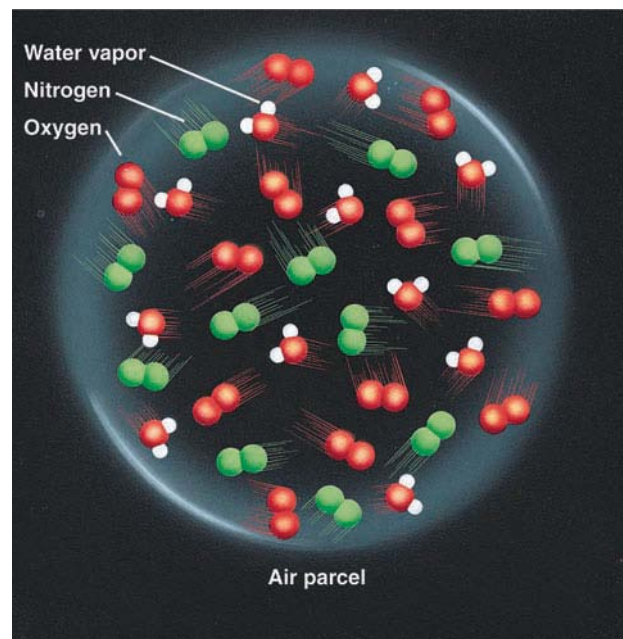
Humidity

We are now ready to look more closely at the concept of **humidity**, which may refer to any one of a number of ways of specifying the amount of water vapor in the air. Since there are several ways to express atmospheric water vapor content, there are several meanings for the concept of humidity. The first type of humidity we’ll take a look at is *absolute humidity*.

ABSOLUTE HUMIDITY Suppose we enclose a volume of air in an imaginary thin elastic container—a *parcel*—about the size of a large balloon, as illustrated in • Fig. 4.6. With a chemical drying agent, we can extract the water vapor from the air, weigh it, and obtain its mass. If we then compare the vapor’s mass with the volume of air in the parcel, we would have determined the **absolute humidity** of the air—that is, the mass of water vapor in a given volume of air, which can be expressed as

$$\text{Absolute humidity} = \frac{\text{mass of water vapor}}{\text{volume of air}}.$$

Absolute humidity represents the *water vapor density* (mass/volume) in the parcel and, normally, is expressed as grams of



• **FIGURE 4.6** The water vapor content (humidity) inside this air parcel can be expressed in a number of ways.

	Parcel Size	Mass of H ₂ O Vapor	Absolute Humidity
	2 m ³	10 g	5 g/m ³
	1 m ³	10 g	10 g/m ³

• **FIGURE 4.7** With the same amount of water vapor in a parcel of air, an increase in volume decreases absolute humidity, whereas a decrease in volume increases absolute humidity.

water vapor in a cubic meter of air. For example, if the water vapor in 1 cubic meter of air weighs 25 grams, the absolute humidity of the air is 25 grams per cubic meter (25 g/m³).

We learned in Chapter 2 that a rising or descending parcel of air will experience a change in its volume because of the changes in surrounding air pressure. Consequently, when a volume of air fluctuates, the absolute humidity changes—even though the air’s vapor content has remained constant (see • Fig. 4.7). For this reason, the absolute humidity is not commonly used in atmospheric studies.

SPECIFIC HUMIDITY AND MIXING RATIO Humidity, however, can be expressed in ways that are not influenced by changes in air volume. When the mass of the water vapor in the air parcel in Fig. 4.6 is compared with the mass of all the air in the parcel (including vapor), the result is called the **specific humidity**; thus

$$\text{Specific humidity} = \frac{\text{mass of water vapor}}{\text{total mass of air}}.$$

Another convenient way to express humidity is to compare the mass of the water vapor in the parcel to the mass of the remaining dry air. Humidity expressed in this manner is called the **mixing ratio**; thus

$$\text{Mixing ratio} = \frac{\text{mass of water vapor}}{\text{mass of dry air}}.$$

Both specific humidity and mixing ratio are expressed as grams of water vapor per kilogram of air (g/kg).

The specific humidity and mixing ratio of an air parcel remain constant *as long as water vapor is not added to or removed from the parcel*. This happens because the total number of molecules (and, hence, the mass of the parcel) remains constant, even as the parcel expands or contracts (see • Fig. 4.8). Since changes in parcel size do not affect specific humidity and mixing ratio, these two concepts are used extensively in the study of the atmosphere.

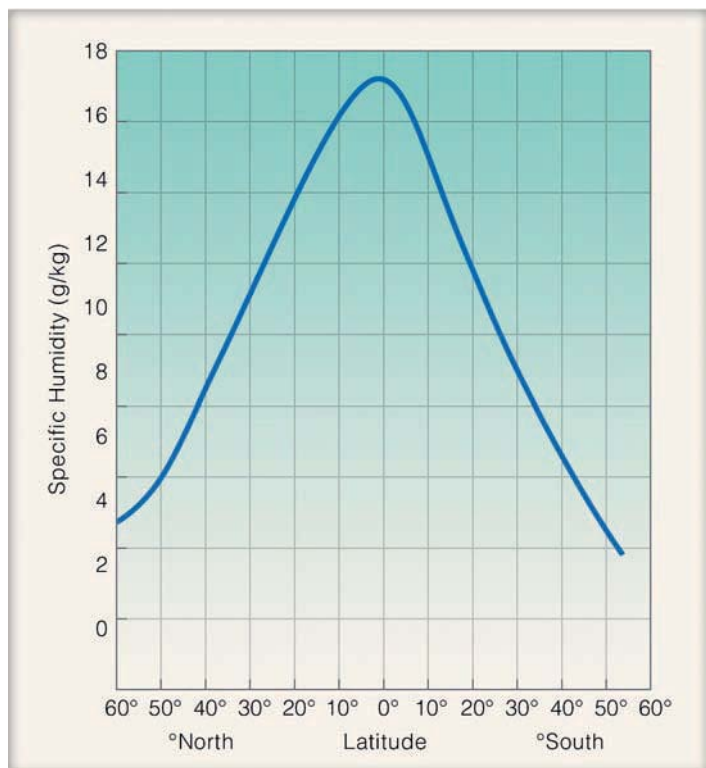
• Figure 4.9 shows how specific humidity varies with latitude. The average specific humidity is highest in the warm, muggy tropics. As we move away from the tropics, it decreases, reaching its lowest average value in the polar latitudes. Although the major deserts of the world are located



● **FIGURE 4.8** The specific humidity does not change as air rises and descends.

near latitude 30°, Fig. 4.9 shows that, at this latitude, the average air contains nearly twice the water vapor than does the air at latitude 50°N. Hence, the air of a desert is certainly not “dry,” nor is the water vapor content extremely low. Since the hot, desert air of the Sahara often contains more water vapor than the cold, polar air farther north, we can say that *summertime Sahara air has a higher specific humidity*. (We will see later in what sense we consider desert air to be “dry.”)

VAPOR PRESSURE The air’s moisture content may also be described by measuring the pressure exerted by the water vapor in the air. Suppose the air parcel in Fig. 4.6, (p. 93), is near sea level. The total pressure inside the parcel is due to the collision of all the molecules against the inside surface of the parcel. In other words, the total pressure inside the parcel is equal to the sum of the pressures of the individual gases. (This phenomenon is known as *Dalton’s law of partial pressure*.) If



● **FIGURE 4.9** The average specific humidity for each latitude. The highest average values are observed in the tropics and the lowest values in polar regions.

the total pressure inside the parcel is 1000 millibars (mb),* and the gases inside include nitrogen (78 percent), oxygen (21 percent), and water vapor (1 percent), then the partial pressure exerted by nitrogen would be 780 mb and by oxygen, 210 mb. The partial pressure of water vapor, called the **actual vapor pressure**, would be only 10 mb (1 percent of 1000).† It is evident, then, that because the number of water vapor molecules in any volume of air is small compared to the total number of air molecules in the volume, the actual vapor pressure is normally a small fraction of the total air pressure.

Everything else being equal, the more air molecules in a parcel, the greater the total air pressure. When you blow up a balloon, you increase its pressure by putting in more air. Similarly, an increase in the number of water vapor molecules will increase the total vapor pressure. Hence, the actual vapor pressure is a fairly good measure of the total amount of water vapor in the air: *High actual vapor pressure indicates large numbers of water vapor molecules, whereas low actual vapor pressure indicates comparatively small numbers of vapor molecules.*‡

In summer across North America, the highest vapor pressures are observed along the humid Gulf Coast, whereas the lowest values are experienced over the drier Great Basin, especially Nevada. In winter, the highest average vapor pressures are again observed along the Gulf Coast with lowest values over the northern Great Plains into Canada.

Actual vapor pressure indicates the air’s total water vapor content, whereas **saturation vapor pressure** describes how much water vapor is necessary to make the air saturated at any given temperature. Put another way, *saturation vapor pressure is the pressure that the water vapor molecules would exert if the air were saturated with vapor at a given temperature.*§

We can obtain a better picture of the concept of saturation vapor pressure by imagining molecules evaporating from a water surface. Look back at Fig. 4.4b, (p. 92) and recall that when the air is saturated, the number of molecules escaping from the water’s surface equals the number returning. Since the number of “fast-moving” molecules increases as the temperature increases, the number of water molecules escaping per second increases also. In order to maintain equilibrium, this situation causes an increase in the number of water vapor molecules in the air above the liquid. Consequently, at higher air temperatures, it takes more water vapor to saturate the air. And more vapor molecules exert a greater pressure. *Saturation vapor pressure, then, depends primarily on the air temperature.* From the graph in ● Fig. 4.10, we can see that at

*You may recall from Chapter 1 that the millibar is the unit of pressure most commonly found on surface weather maps, and that it expresses atmospheric pressure as a force over a given area.

†When we use the percentages of various gases in a volume of air, Dalton’s law only gives us an approximation of the actual vapor pressure. The point here is that, near the earth’s surface, the actual vapor pressure is often close to 10 mb.

‡Remember that actual vapor pressure is only an approximation of the total vapor content. A change in total air pressure will affect the actual vapor pressure even though the total amount of water vapor in the air remains the same.

§When the air is saturated, the amount of water vapor is the maximum possible at the existing temperature and pressure.

10°C, the saturation vapor pressure is about 12 mb, whereas at 30°C it is about 42 mb.

The insert in Fig. 4.10 shows that, when both water and ice exist at the same temperature below freezing, the saturation vapor pressure just above the water is greater than the saturation vapor pressure over the ice. In other words, at any temperature below freezing, it takes more vapor molecules to saturate air directly above water than it does to saturate air directly above ice. This situation occurs because it is harder for molecules to escape an ice surface than a water surface. Consequently, fewer molecules escape the ice surface at a given temperature, requiring fewer in the vapor phase to maintain equilibrium. Likewise, salts in solution bind water molecules, reducing the number escaping. These concepts are important and (as we will see in Chapter 7) play a role in the process of rain formation.

So far, we've described the amount of moisture actually in the air. If we want to report the moisture content of the air around us, we have several options:

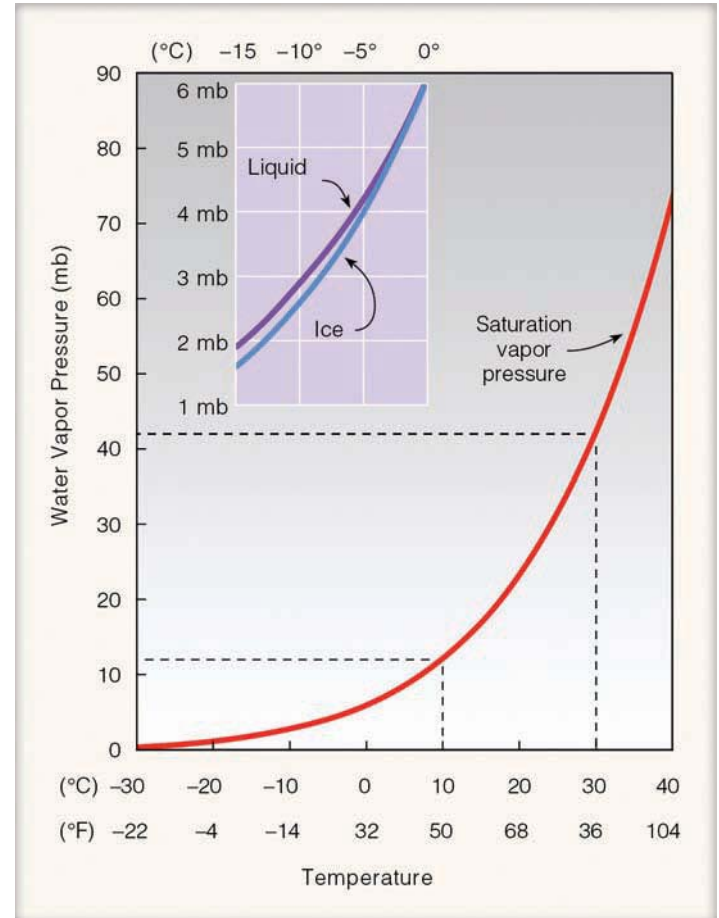
1. *Absolute humidity* tells us the *mass* of water vapor in a fixed volume of air, or the *water vapor density*.
2. *Specific humidity* measures the *mass* of water vapor in a fixed *total mass* of air, and the *mixing ratio* describes the mass of water vapor in a fixed mass of the remaining dry air.
3. The *actual vapor pressure* of air expresses the amount of water vapor in terms of the amount of *pressure* that the water vapor molecules exert.
4. The *saturation vapor pressure* is the pressure that the water vapor molecules would exert if the air were saturated with vapor at a given temperature.

Each of these measures has its uses but, as we will see, the concepts of vapor pressure and saturation vapor pressure are critical to an understanding of the sections that follow. (Before looking at the most commonly used moisture variable—relative humidity—you may wish to read the Focus section on vapor pressure and boiling, p. 96.)

RELATIVE HUMIDITY While relative humidity is the most common way of describing atmospheric moisture, it is also, unfortunately, the most misunderstood. The concept of relative humidity may at first seem confusing because it does not indicate the actual amount of water vapor in the air. Instead, it tells us how close the air is to being saturated. The **relative humidity** (RH) is the ratio of the amount of water vapor actually in the air to the maximum amount of water vapor required for saturation at that particular temperature (and pressure). It is the ratio of the air's water vapor content to its capacity; thus

$$\text{RH} = \frac{\text{water vapor content}}{\text{water vapor capacity}}.$$

We can think of the actual vapor pressure as a measure of the air's actual water vapor content, and the saturation vapor



ACTIVE FIGURE 4.10 Saturation vapor pressure increases with increasing temperature. At a temperature of 10°C, the saturation vapor pressure is about 12 mb, whereas at 30°C it is about 42 mb. The insert illustrates that the saturation vapor pressure over water is greater than the saturation vapor pressure over ice. Visit the Meteorology Resource Center to view this and other active figures at academic.cengage.com/login

pressure as a measure of air's total capacity for water vapor. Hence, the relative humidity can be expressed as

$$\text{RH} = \frac{\text{actual vapor pressure}}{\text{saturation vapor pressure}} \times 100 \text{ percent.}^*$$

Relative humidity is given as a percent. Air with a 50 percent relative humidity actually contains one-half the amount required for saturation. Air with a 100 percent relative humidity is said to be *saturated* because it is filled to capacity with water vapor. Air with a relative humidity greater than 100 percent is said to be **supersaturated**. Since relative humidity is used so much in the everyday world, let's examine it more closely.

*Relative humidity may also be expressed as

$$\text{RH} = \frac{\text{actual mixing ratio}}{\text{saturation mixing ratio}} \times 100 \text{ percent,}$$

where the actual mixing ratio is the mixing ratio of the air, and the saturation mixing ratio is the mixing ratio of saturated air at that particular temperature.