4.3.2 Synoptic-Scale Air Circulation

Air Masses

At the *synoptic scale,* which corresponds to roughly a thousand kilometers, the formation and migration of tropospheric *air masses* are responsible for many of the day-to-day changes in weather and air quality. Air masses form when air remains over a large, fairly homogeneous region of Earth's surface, such as a desert or tropical ocean, for a sufficient period of time to acquire distinctive characteristics, especially temperature and moisture content, from the surface. Terrestrial and oceanic regions give rise to *continental* and *maritime* air masses, respectively. A continental air mass usually has a lower moisture content, and often a more extreme temperature (hot or cold, depending on its latitude of origin), than a maritime air mass; a maritime air mass generally is closer to moisture saturation than a continental air mass. Polar air masses are relatively colder, while tropical air masses are relatively warmer, than other air masses that they encounter.

The migration of an air mass results in transport of its associated temperature, moisture, and chemical contents. Along boundaries between air masses of differing temperature and moisture content, called *fronts,* the warmer, less

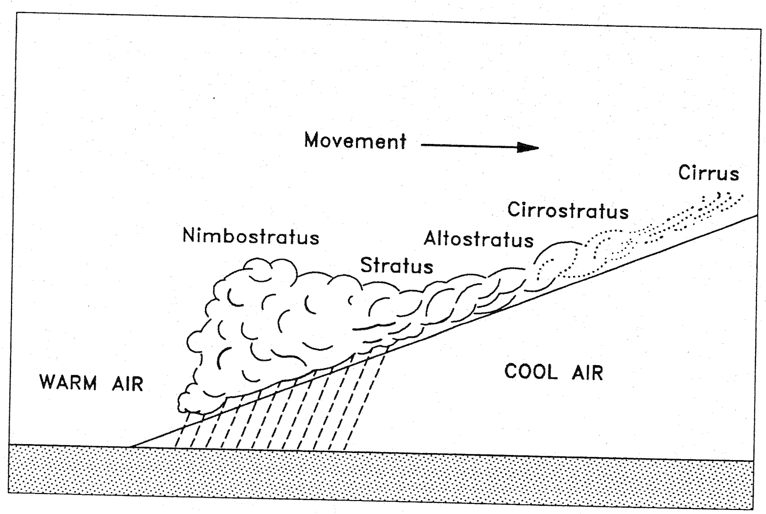

FIGURE 4-15a A warm front occurs when a warm air mass overtakes an adjoining cold air mass and rises over it. Adiabatic cooling tends to produce condensation, and ultimately precipitation, along the front. In the situation shown, the warm air mass is unconditionally stable. If the warm air mass is conditionally stable, instability is produced when clouds form along the front, and vertical cloud development (often including thunderstorms) occurs.

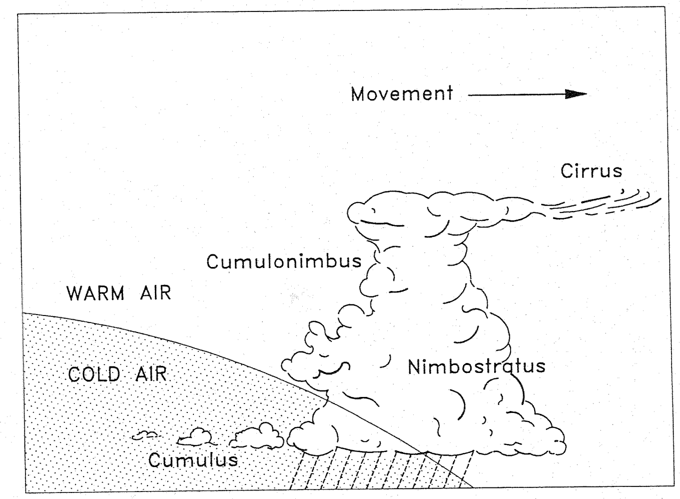

FIGURE 4-15b A cold front occurs when a cold air mass overtakes a warm air mass, which it displaces upward. Again, adiabatic cooling of the rising air tends to produce precipitation. Cold fronts are usually faster moving than are warm fronts; they are also narrower, with the slope of the interface between air masses typically about 80 : 1.

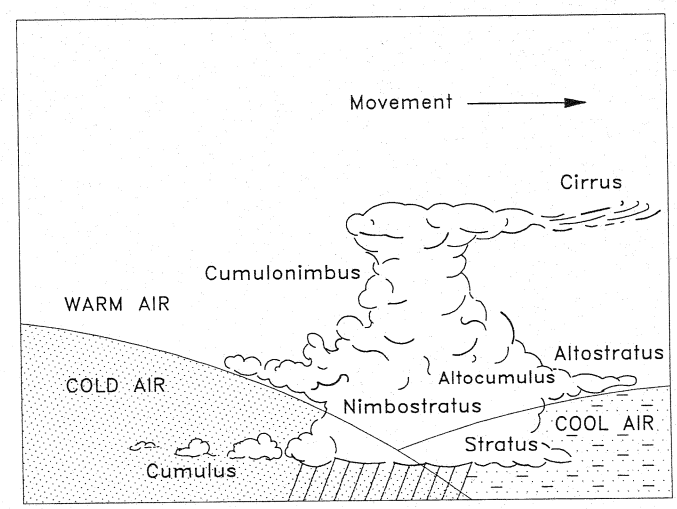

FIGURE 4-15c An occluded front occurs when a warm air mass rises under the combined influence of two colder air masses, causing extensive cloud development and precipitation.

dense air mass tends to rise over the cooler, denser air; because of adiabatic cooling, this frequently results in precipitation. When a warm air mass overtakes a cold air mass, the front is called a *warm front* (Fig. 4-15a); when a cold air mass overtakes a warm air mass, a *cold front* results (Fig. 4-15b). A cold front is usually narrower and faster moving than a warm front, but both can produce clouds and precipitation. An *occluded front* results when a cold front catches up to a more slowly moving warm front, and warmer air becomes trapped between two colder air masses (Fig. 4-15c).

Cloud patterns associated with fronts frequently can be seen from the perspective of a satellite, but a person on the ground, who cannot observe a front in its entirety, must infer its presence from the characteristic progressions of cloud types and weather that occur.

Cloud types often are classified based on altitude. *High clouds* have their bases above 7 km (23,000 ft) and include the wispy mare's tail clouds known as *cirrus;* the *cirrocumulus,* known as "mackerel sky"; and the layers of *cirrostratus. Middle clouds* have altitudes between 2 and 7 km (6500 to 23,000 ft), and are either the rounded *altocumulus* or the layered *altostratus. Low clouds* have bases from near Earth's surface to about 2 km (6500 ft), and include *stratocumulus, stratus,* and *nimbostratus.* Nimbostratus clouds usually bring rain or snow. Clouds with vertical development extend from about 2 to 7 km or more, and include *cumulonimbus* (thunderhead clouds) and *cumulus.*

A passing cold front is heralded by clouds, a drop in temperature, and precipitation; cooler air and clear skies occur behind the cold front (see Fig. 4-15b). The slower moving warm front is characterized by a more gradual lowering of cloud heights, followed by rain or snow (Fig. 4-15a). As air masses pass, so do their burdens of airborne chemicals. The clouds and precipitation formed along the front act as sinks for certain atmospheric chemicals because of rainout and washout processes. These processes, which remove particles, gases, and dissolved chemicals from the atmosphere and deposit them on Earth's surface, are discussed in Section 4.5.

Cyclones and Anticyclones

While some aspects of weather can be explained in terms of air mass movements, other synoptic-scale features of weather systems are associated with *cyclones* and *anticyclones.* Cyclones and anticyclones are large eddies (hundreds of kilometers across) in the atmosphere. Within their boundaries they influence and often control the direction and speed of wind and dominate the weather on a regional scale. They also influence the advective transport of air pollutants and subsequent air quality. On a global scale, cyclones and anticyclones contribute to tropospheric mixing.

Contrary to lay terminology and certain regional usages, a cyclone does not necessarily connote a violent windstorm. In its most general usage, a cyclone is an eddy whose direction of rotation is the same as that of the hemisphere within which it lies (i.e., counterclockwise as viewed from above in the Northern Hemisphere, clockwise in the Southern Hemisphere).

The characteristics of air circulation in a cyclone can be explained in terms of four forces: air pressure gradient, the Coriolis force, friction, and centrifugal force. Two of these forces, the Coriolis force and centrifugal force, are so-called virtual forces that exist because of the acceleration of the reference system. These two forces are treated differently in a nonaccelerating, or *inertial,* reference system. (To an observer in an inertial reference system, the stars appear motionless.) In the following discussion the chosen reference frame is an air parcel that is part of a cyclone or an anticyclone; this reference frame has acceleration due to both the rotation of Earth (giving rise to the Coriolis force) and the circular movement of air within the cyclone or the anticyclone (giving rise to centrifugal force). Within this reference system the sum of all forces is zero. (Note that if an inertial reference frame is chosen, the sum of all forces acting on a parcel moving at velocity v with radius R must have an acceleration of magnitude v^2/R, directed toward the center of the curved path.)

Thus, within the reference frame of a parcel of air circulating about the center of a cyclone, the air parcel experiences the forces shown in Fig. 4-16. The Coriolis force acts radially outward, in a direction perpendicular to, and to the right of, the velocity of the reference system. The centrifugal force also acts radially outward. If the air parcel is located higher than a few hundred meters above Earth's surface, friction may be neglected in an approximate analysis. When friction is neglected, the sum of the Coriolis force and the centrifugal force is balanced by an air pressure force that acts in an inward direction due to the air pressure being lower in the core of the cyclone than outside the cyclone.

In many cyclones, the Coriolis force is larger than the centrifugal force. However, because the centrifugal force is proportional to v^2/R while the Coriolis force is proportional to v, the centrifugal force increases more rapidly than the Coriolis force as a cyclone becomes small and/or its velocity increases. Thus, in a *hurricane* (a violent tropical cyclone over the Atlantic Ocean), the centrifugal force may substantially exceed the Coriolis force. In a *tornado* (a localized, violent midlatitude storm usually originating over land), the Coriolis force may be negligible compared with the centrifugal force because a tornado is smaller and has even higher wind speeds than a hurricane. As testimony to the dominance of centrifugal force under such conditions, *clockwise* tornadoes are sometimes observed in storm systems in the Northern Hemisphere.

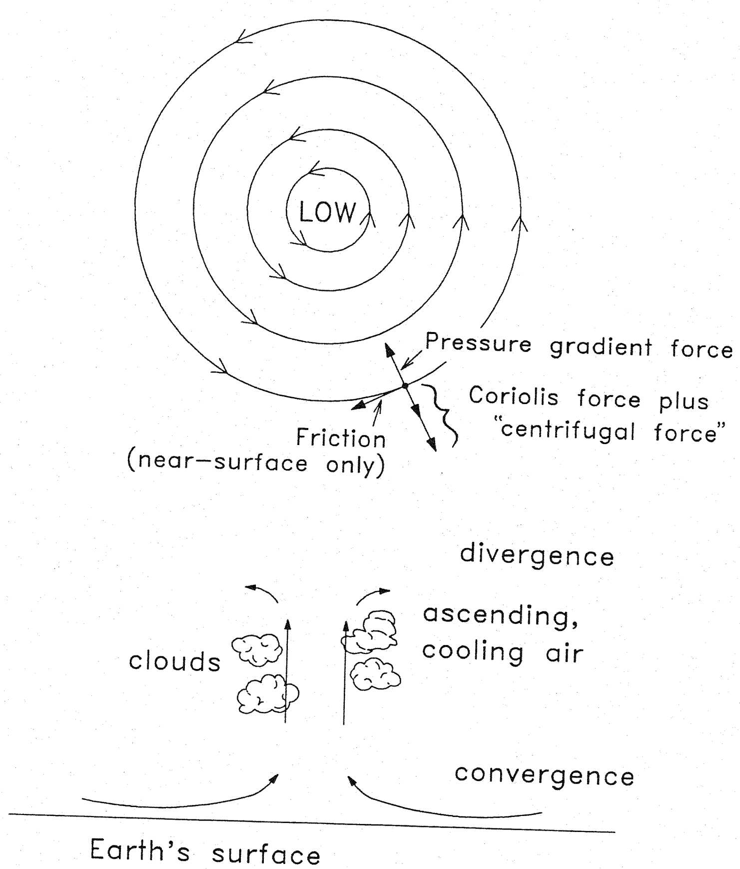

FIGURE 4-16 Forces acting on air in a cyclone. Wind is counterclockwise as viewed from above in the Northern Hemisphere; therefore, both the Coriolis force and centrifugal force are outward. Their sum is balanced by the pressure gradient force. At low altitude, where friction with the ground causes the air velocity to decrease, the Coriolis force and centrifugal force are correspondingly weakened, while the pressure gradient force is not; this results in low-level air being drawn into the core, where it then must ascend to satisfy mass balance constraints. Condensation and precipitation are a likely result in the center of the cyclone.

The low pressure at the center of a cyclone results in large cyclones often being called *lows* or *low pressure areas*. Near the ground surface, frictional forces retard the movement of air, decreasing its velocity and therefore lessening the Coriolis force, Eq. [4-13]. Near-surface air slowed by friction is drawn by the air pressure gradient toward the center of the cyclone. This *convergence* of surface air must, by mass balance considerations, be accompanied by an upward flow of air through the core of the cyclone.

As air is lifted through the center of a cyclone, it often reaches its dew point, forming clouds and creating precipitation. If sufficiently warm and moisture-laden air is drawn into a cyclone, the heat energy released by condensation can augment the kinetic energy of the cyclone, causing it to intensify. This process occurs in hurricanes over warm, tropical seas.

The balance of forces on a parcel of air moving in an anticyclone is shown in Fig. 4-17. The reference system is the accelerating reference system of the air parcel itself. In this case, the Coriolis force acts radially inward due to the clockwise movement of the air. This inward force is balanced by the outward centrifugal force and the outward pressure gradient. Because the anticyclone has higher pressure at its core than outside, it is commonly called a *high* or *high-pressure area*. Friction causes the air velocity near Earth's surface to be lower than the velocity aloft; this in turn decreases the Coriolis force, allowing near-surface air to flow outward from the high-pressure center of the anticyclone.

Conservation of mass requires air to descend within the core of the anticyclone to replace that lost by outflow near Earth's surface. Adiabatic warming occurs as the air descends; precipitation in an anticyclone is therefore suppressed, and high-pressure areas are generally associated with fair weather. Anticyclones may, however, be associated with episodes of lowered air quality, in part because the downward movement of air can create atmospheric inversions, causing pollutants to build up to elevated concentration levels in urban areas. This air pollutant buildup is further aggravated by the light winds of high-pressure areas that lead to low rates of pollutant removal (ventilation). Although winds in a low-pressure area can reach very high velocities, because arbitrarily large pressure gradients can be balanced by arbitrarily large centrifugal forces associated with high wind speeds, winds in a high-pressure area are constrained to be light. In an anticyclone, a finite Coriolis force is balanced by both the pressure gradient and the centrifugal force; increasingly higher pressures cannot be balanced by increasingly higher wind speeds, because the outward centrifugal force also would increase rapidly (proportional to v^2) at higher wind speeds.

Three of the most acutely health-threatening episodes of air pollution in the 20th century were associated with high-pressure areas. These notorious

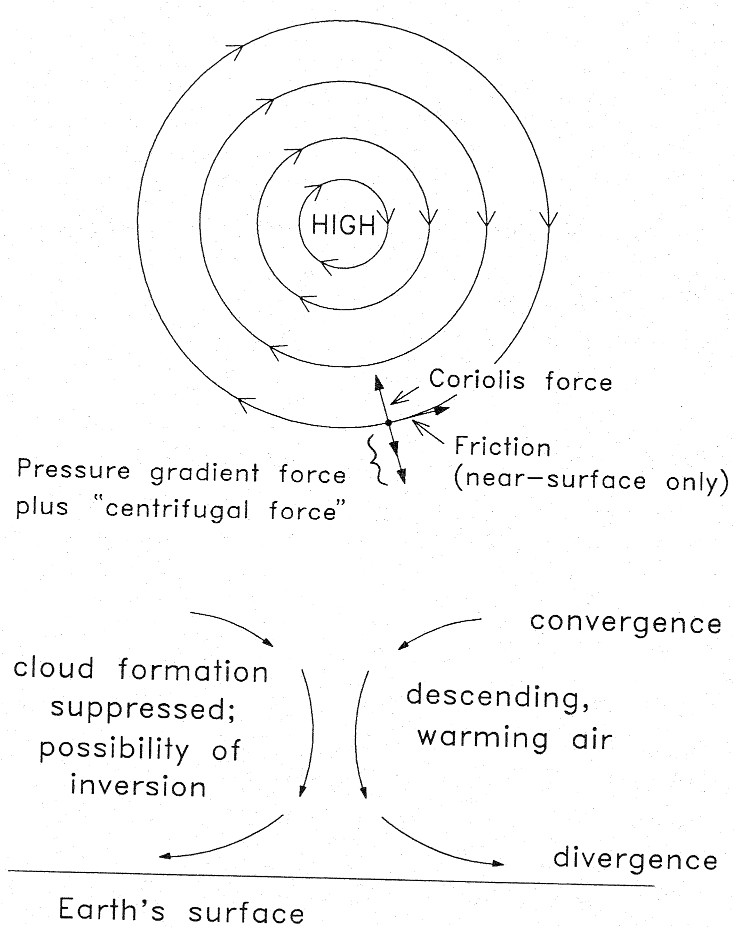

FIGURE 4-17 Forces acting on air in an anticyclone. In the Northern Hemisphere, the Coriolis force acts toward the center of the anticyclone, balanced by the outward centrifugal force and the outward pressure gradient force. Near ground level, Coriolis force and centrifugal force are both weaker, because the air velocity is slowed by friction with the ground; air flows outward near ground level under the influence of the pressure gradient force and therefore descends within the core of the anticyclone. Descending air is warmed adiabatically, inhibiting precipitation and sometimes creating an inversion.

episodes occurred in the Meuse Valley, Belgium, December 1 to 5, 1930; in Donora, Pennsylvania, October 25 to 31, 1948; and in London, United Kingdom, December 5 to 9, 1952. In these cases, strong atmospheric inversions associated with anticyclones helped trap pollutants such as SO_2, resulting in increased air pollutant concentrations. These elevated concentrations caused elevated rates of illness and mortality (Wilson and Spengler, 1996).

Relationships between Frontal Masses and Cyclones

Frequently, weather systems involve a close interaction among fronts, cyclones, and anticyclones. The weather system shown in Fig. 4-16 has a large cyclone (low-pressure area) located at 180° west. A warm front, indicated by a line with semicircles pointing in the direction of movement, extends eastward from the cyclone, while a cold front, indicated by a line with triangles pointing in the direction of movement, extends southward. An occluded front extends a short distance northward. Note that clouds and precipitation are associated with this system, especially to the north of the low-pressure area and along the fronts.

The spatial relationships among fronts and cyclones, such as seen in Fig. 4-18, are frequently more than just coincidental; they are often causal. Mid-

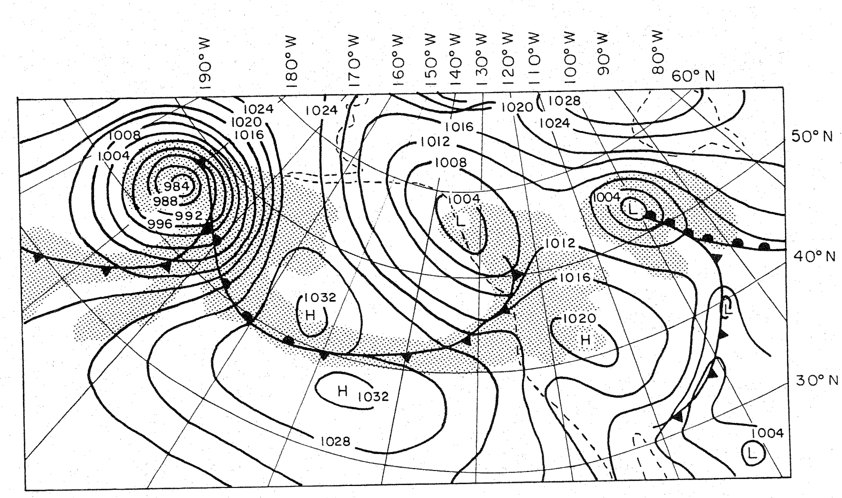

FIGURE 4-18 Tracing of a satellite photo showing cloud formation along fronts, as well as the relationship between low-pressure areas (cyclones) and fronts. Note the correspondence between the cyclones and fronts in this figure and the idealized patterns shown later in Fig. 4-20 for the genesis of a cyclone. Pressures are in millibars; dotted lines show the west coast of North America and Hudson Bay (adapted from FAA, 1965).

latitude cyclones can be formed by *shear* (motion parallel to the front) between the two air masses constituting a front. It is a general principle that shear in fluids tends to produce eddies; familiar small-scale examples include vortices behind the oars of a row boat, or behind a spoon while stirring a cup of coffee. In rivers, the turbulence that gives rise to mixing is generated in part by shear between the more slowly moving water near the river bottom and the faster moving water above it. In the ocean, temperature differences seen by satellite reveal eddies (*warm* and *cold rings*) created by shear between

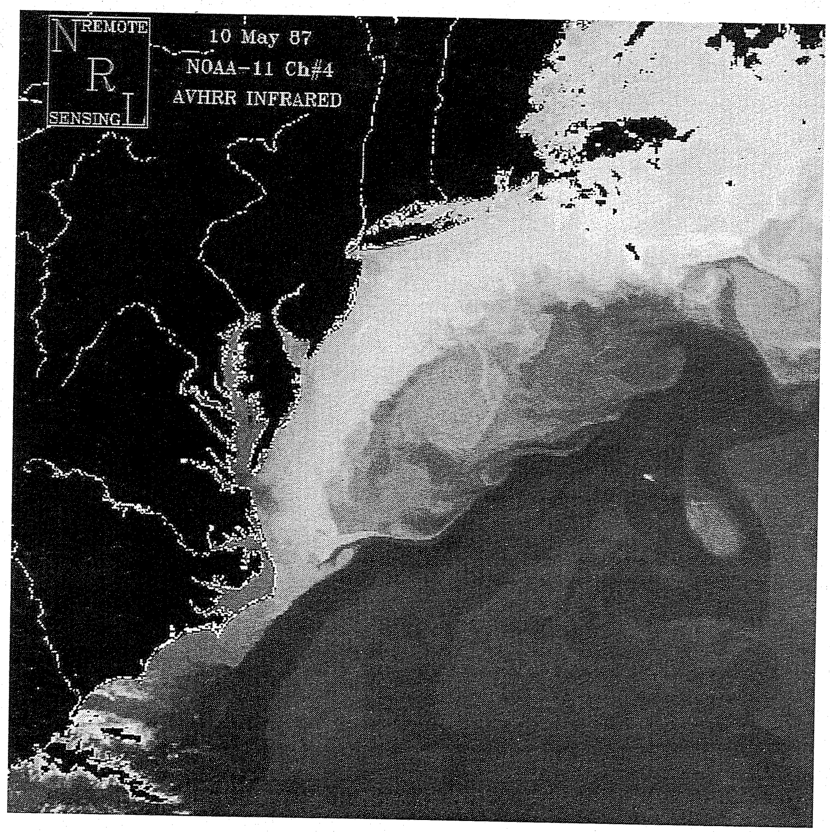

FIGURE 4-19 May 10, 1987 imagery of eddies along the boundary between the warmer (darker gray) Gulf Stream and colder (lighter gray) North Atlantic water. This large-scale eddy generation is similar to the generation of cyclones along the polar front, where shear between air masses occurs (courtesy J. Hawkins and D. May, Naval Remote Sensing Laboratory).

the warm northward-moving Gulf Stream and adjoining cooler waters of the North Atlantic (Fig. 4-19). Although less readily visible, the same process of eddy production by shear occurs in the atmosphere. In the Northern Hemisphere, a regular progression of low-pressure systems is spawned along the interface between the midlatitude westerlies and the colder polar air masses with their prevailing easterly winds. Steps in the genesis of a cyclone along this *polar front* are shown in Fig. 4-20.

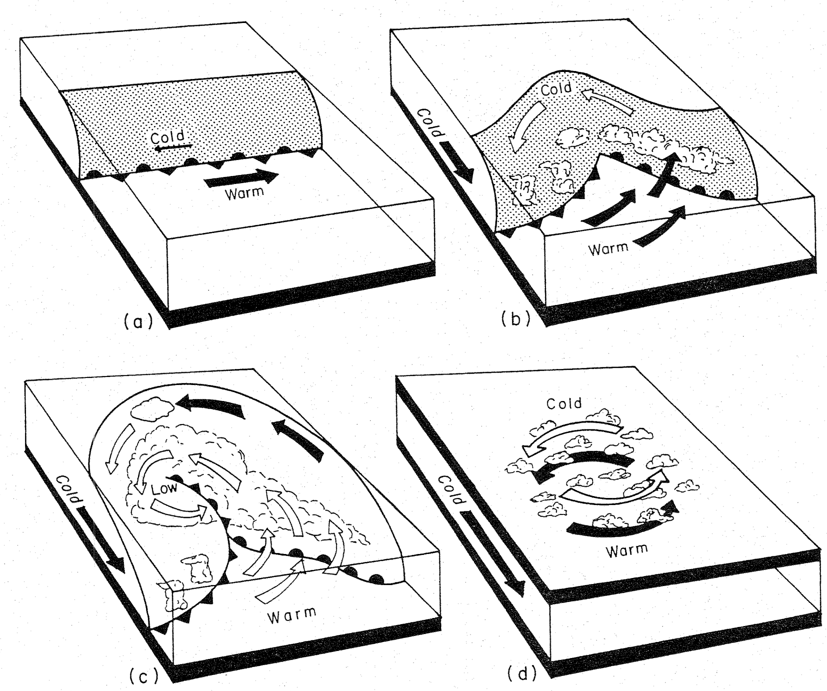

FIGURE 4-20 Formation of a midlatitude cyclone along the polar front. (a) Initially the air masses form a stationary front. Although the front is not moving, the prevailing winds cause shear along the front. (b) The cyclone (low-pressure area) begins as the warm midlatitude air at one location moves poleward over the polar air, creating a warm front; elsewhere the polar air moves southward, creating a cold front. The beginnings of cyclonic air circulation are evident. (c) In a mature system, part of the faster moving cold front may catch up with the warm front, creating an occluded front. Note the similarity of this arrangement of fronts to the weather system of Fig. 4-18. (d) In this dissipated stage, mixing between the air masses has caused the fronts to lose their identities. Moisture has been removed from the system as precipitation. (Adapted from Frederick K. Lutgens/Edward J. Tarbuck, *The Atmosphere: An Introduction to Meteorology*, 5th ed., © 1992, p. 217. Reprinted by permission of Prentice Hall, Englewood Cliffs, New Jersey.)

4.3.3 The Synoptic Weather Map

Because of the great importance of weather to many endeavors, including agriculture, ship navigation, surface travel, and aviation, extensive weather reporting and forecasting have been implemented over large areas of the globe. Over the oceans, where surface-based meteorological data are sparse, satellite imagery has enhanced coverage of weather phenomena. Synoptic-

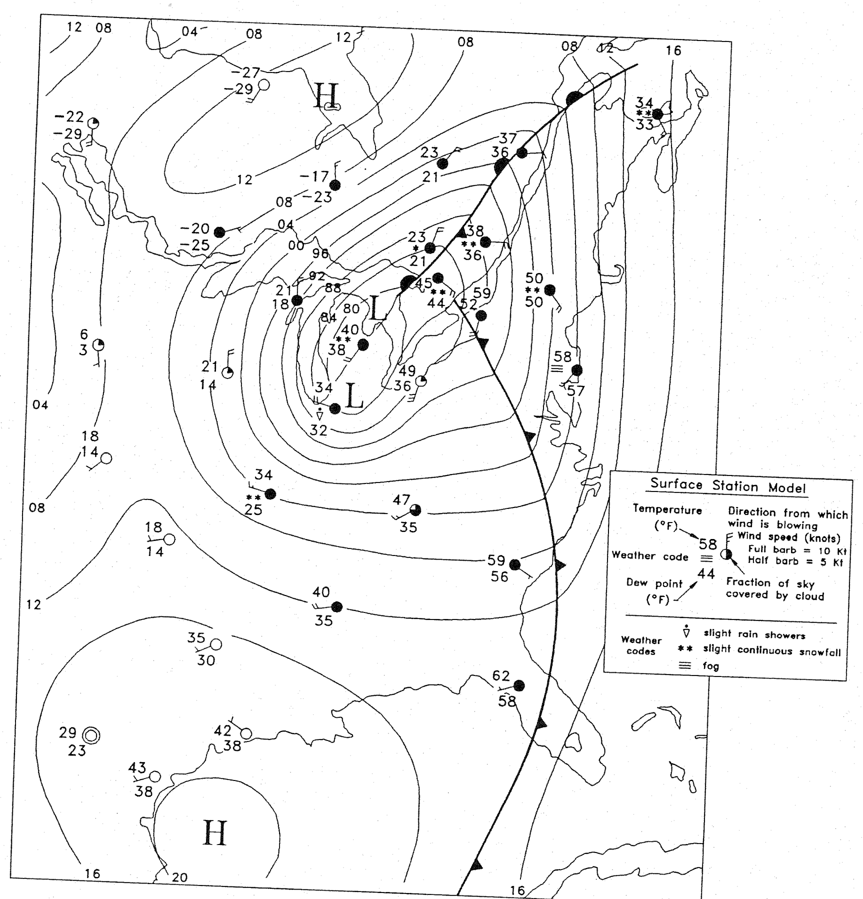

FIGURE 4-21 A surface weather chart for 1200 GMT (700 EST), March 5, 1964, showing information available at each observation station. The chart also shows high- and low-pressure areas, fronts, and isobars constructed by interpolation between stations. (900 or 1000 millibars need to be added as appropriate to values; see text. Adapted from FAA, 1965).

scale weather data are commonly summarized in *surface weather charts*, such as that shown in Fig. 4-21.

Cloud cover, surface wind direction and speed, temperature, dew point, and precipitation are indicated by symbols at each weather reporting station. Interpretation of the weather in a region is facilitated by the identification of high-pressure and low-pressure areas, the delineation of frontal systems, and the construction of *isobars*. Isobars are lines of constant atmospheric pressure that are interpolated between observation points, and are so named because pressure often is presented using the unit of the bar, which is equivalent to 10^6 dyn/cm^2 or 0.9869 atm. On some maps, such as shown in Fig. 4-21, pressure is given in millibars; as a shorthand notation, only the last two digits are presented. The user must add either 900 or 1000 millibars to the value shown, whichever brings the value closest to 1 bar.

Figure 4-21 shows a major low-pressure system over the Great Lakes and the eastern coast of the United States, with a cold front located to the southeast and an occluded front to the northeast. In this late winter storm system, surface winds are counterclockwise and blowing somewhat toward the center of the low, as would be expected from Fig. 4-16. By convention, the tails of the wind "arrows" extend in the direction from which the wind is blowing, and the barbs on the arrow tails indicate wind speed, which is generally greater for stations closer to the center of the low. The various precipitation symbols, and the filled-in or partially filled-in station circles that reflect the fraction of sky covered by clouds, show that precipitation and clouds are associated with this system, especially near the fronts. Southward, behind the cold front, clear skies and cooler temperatures prevail; RH also is lower, as shown by the larger differences between temperature and dew point.

The data presented in weather maps are relevant to chemical transport both at the local scale, where local weather governs atmospheric mixing, and at the regional scale, where synoptic-scale circulation governs the long-range transport of chemicals across state and national boundaries.

4.3.4 Local Effects

At scales smaller than the synoptic scale, topography and differences in the surface cover (e.g., forest, agricultural fields, open water, or urbanized land) influence local winds, precipitation, and temperature. One common example of a local effect due to surface cover is the *sea breeze*, which occurs because water bodies warm and cool more slowly than the land does. During the day in coastal areas, air over the land warms and rises more rapidly and is replaced by cooler air originating from over the water. The reverse may happen at night, as the land cools to a temperature lower than that of the water body,

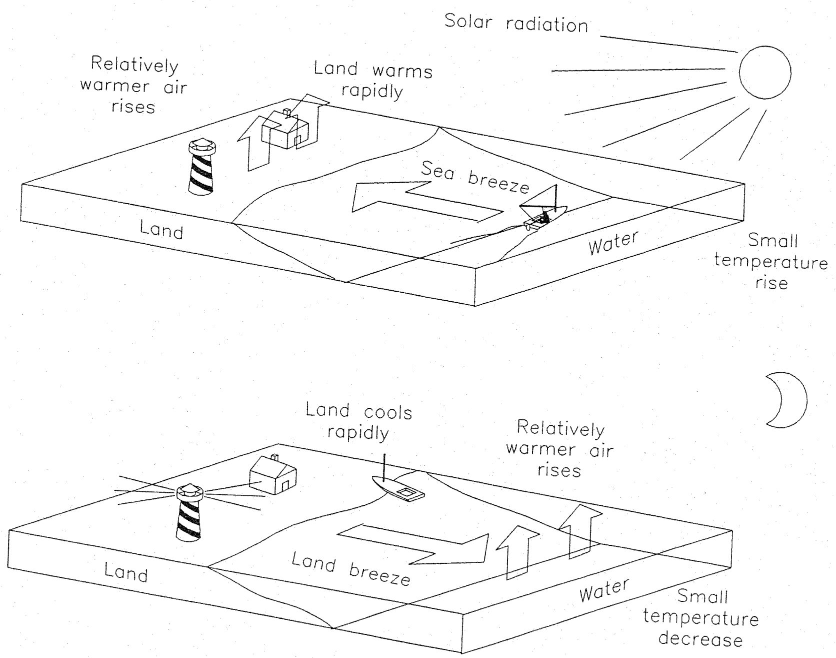

FIGURE 4-22a Sea and land breezes. The sea breeze is a local effect, caused by the faster heating of a land surface than a large water body nearby. The water body temperature rises more slowly than that of the land due to a higher heat capacity and the downward mixing of heat in the water. Sea breezes typically begin in midmorning as air rising over the land is replaced by cooler air from the lake or ocean. At night, when the land becomes cooler than the water, a land breeze may be established.

and air cooled by the land flows toward the water (Fig. 4-22a). Because of the great heat capacity of a large water body, coastal areas also are likely to experience more moderate temperature fluctuations throughout the year than are inland areas. In addition, areas located downwind of large water bodies may receive more precipitation than surrounding areas (the *lake effect*), because the water bodies supply moisture to the air more rapidly than does the land surface.

Topography also influences local winds, precipitation, and temperature. For example, *valley winds* can form as air along a hillside is warmed during the day and thus rises, drawing air up the valley, or as air cools at night and thus flows downhill toward the valley bottom (Fig. 4-22b). In another topographic effect, the windward side of a mountain range often receives extra

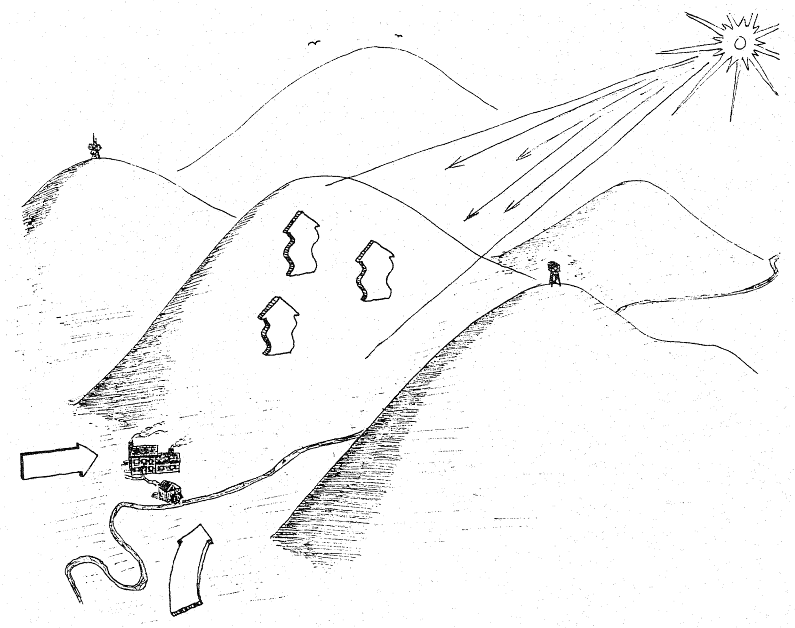

FIGURE 4-22b Valley winds. Like the sea breeze, a valley wind is created by uneven solar heating of Earth's surface. In response to heating by the Sun, air rises along the sides of the valley, drawing air up the valley. Under nighttime conditions, cold air may drain down the valley in response to rapid cooling along the hillsides.

precipitation due to orographic uplifting and associated adiabatic cooling (partially along a wet adiabat) of humid air. The leeward side of such a range tends to be drier and warmer as air, which already has lost moisture on the windward side, descends and warms adiabatically (along a dry adiabat). This *rain shadow* effect, shown in Fig. 4-22c, is prominent along the coastal mountain ranges of the northwestern United States.

Another local effect, in this case due to surface cover, is the urban heat island. Rapid heating of urban pavements and buildings occurs during the daytime because of the high absorbance of constructed surfaces and the absence of cooling from evapotranspiration. At night, rural areas cool more effectively than urban areas because of the relatively unobstructed exposure of the land surface, in contrast to the impediment to heat radiation presented by tall, closely spaced buildings. As a result, daily minimum and maximum

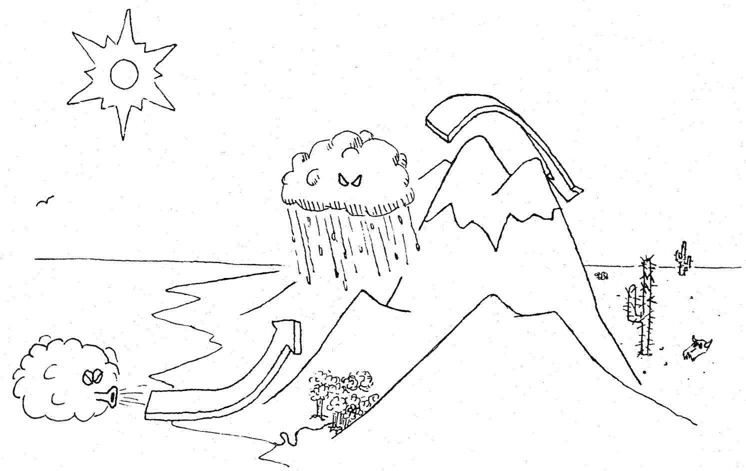

FIGURE 4-22c The rain shadow effect. Moist air, shown here as coming off the ocean, is forced to rise by the presence of a mountain range (orographic lifting). As it rises, it cools according to the dry adiabatic lapse rate, until the dew point is reached. Precipitation then can occur as the air continues to rise along the windward side of the mountain range, its further cooling corresponding to a wet adiabatic lapse rate. The air loses much of its original moisture to precipitation, and as it descends on the leeward side of the mountains, it warms according to the dry adiabatic lapse rate. Thus, when the air has descended on the leeward side to its original altitude (sea level in this example), it is drier and warmer than it was on the windward side. The result is a drier, warmer climate and a dearth of precipitation on the leeward side.

temperatures in urban areas are often higher than those found in outlying areas.

For further information on meteorological processes, the reader is referred to Houghton (1985), Oke (1978), Miller and Thompson (1975), Lutgens and Tarbuck (1992), and Hess (1959).

EXAMPLE 4-4

Humid, fog-laden air at a temperature of 65°F from over the Pacific Ocean is deflected up over a 4000-ft coastal mountain. On the inland side of the mountain, the air follows the land contours and descends to a valley 1000 ft above sea level. What is the temperature and humidity of the air in the valley?

Assume that the initial RH of the air is 100% because of the fog. Use Fig. 4-7, begin at 65°F and sea level (0 m), and follow the slope of the wet adiabat

until the air parcel reaches 4000 ft (approximately 1200 m). At that altitude, the temperature is approximately 54°F. Because the air is still water saturated, it has a partial pressure of water vapor of approximately 11 mm Hg (see Table 4-3). Descending, the air warms along a dry adiabat to approximately 72°F at 1000 ft. The equilibrium vapor pressure of water at 72°F is approximately 20 mm Hg; thus, the new RH (neglecting any pressure correction) is approximately

$$11/20 \approx 55\%.$$

For increased accuracy, a correction for pressure change can be made. Note that the number of moles of water vapor per mole of air (the mixing ratio of water vapor) must be constant during the descent of the air. However, as the pressure of the air increases, the partial pressure of water vapor increases in the same proportion. From Table 4-1, the air pressure at 4000 ft is approximately 0.89 atm; at 1000 ft, it is approximately 0.97 atm. By correcting the partial pressure of water vapor for the compression during descent,

$$11 \text{ mm Hg} \cdot 0.97/0.89 \approx 12 \text{ mm Hg}.$$

The RH, therefore, is more accurately calculated as

$$12 \text{ mm Hg}/20 \text{ mm Hg} \approx 60\%.$$
