5

The Physics of the Atmosphere

The atmosphere serves as the medium through which air pollutants are transported and dispersed. While being transported, the pollutants may undergo chemical reactions and, in addition to removal by chemical transformations, may be removed by physical processes such as gravitational settling, impaction, and wet removal.

This chapter provides an introduction to basic concepts of meteorology necessary to an understanding of air pollution meteorology without specific regard to air pollution problems.

I. ENERGY

All of the energy that drives the atmosphere is derived from a minor star in the universe—our sun. The planet that we inhabit, earth, is 150 million km from the sun. The energy received from the sun is radiant energy—electromagnetic radiation (discussed in Chapter 7). The electromagnetic spectrum is shown in Fig. 5.1. Although this energy is, in part, furnished to the atmosphere, it is primarily received at the earth's surface and redistributed by several processes.

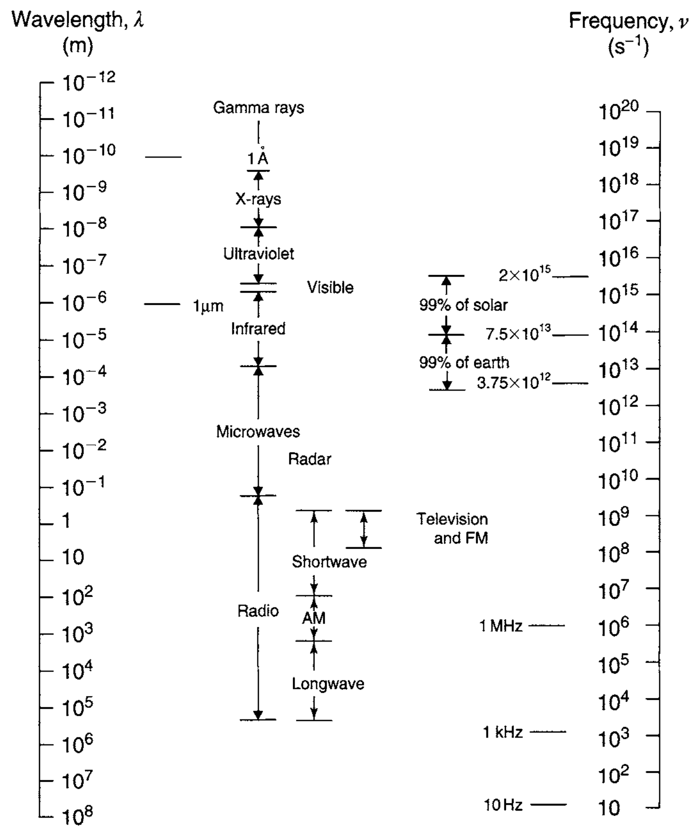

Fig. 5.1. Electromagnetic spectrum. Note the regions of solar and earth radiation.

The earth's gravity keeps the thin layer of gases that constitute the atmosphere from escaping. The combination of solar heating and the spin of the earth causes internal pressure forces in the atmosphere, resulting in numerous atmospheric motions. The strength of the sun's radiation, the distance of the earth from the sun, the mass and diameter of the earth, and the existence and composition of the atmosphere combine to make the earth habitable. This particular combination of conditions would not be expected to occur frequently throughout the universe.

As noted in Chapter 2, the atmosphere is approximately 76% nitrogen, 20% oxygen, 3% water, 0.9% argon, and 0.03% carbon dioxide; the rest consists of relatively inert gases such as neon, helium, methane, krypton, nitrous oxide, hydrogen, and xenon. Compared with the average radius of the earth, 6370 km, the atmosphere is an incredibly thin veil; 90% is below 12 km and 99% below 30 km. In spite of its thinness, however, the total mass of the atmosphere is about 5×10^{18} kg. Therefore, its heat content and energy potential are very large.

A. Radiation from a Blackbody

Blackbody is the term used in physics for an object that is a perfect emitter and absorber of radiation at all wavelengths. Although no such object exists in nature, the properties describable by theory are useful for comparison with materials found in the real world. The amount of radiation, or radiant flux over all wavelengths (F), from a unit area of a blackbody is dependent on the temperature of that body and is given by the Stefan–Boltzmann law:

$$F = \sigma T^4 \tag{5.1}$$

where σ is the Stefan–Boltzmann constant and equals 8.17×10^{-11} cal cm^{-2} min^{-1} deg^{-4} and T is the temperature in degrees K. Radiation from a blackbody ceases at a temperature of absolute zero, 0 K.

In comparing the radiative properties of materials to those of a blackbody, the terms *absorptivity* and *emissivity* are used. Absorptivity is the amount of radiant energy absorbed as a fraction of the total amount that falls on the object. Absorptivity depends on both frequency and temperature; for a blackbody it is 1. Emissivity is the ratio of the energy emitted by an object to that of a blackbody at the same temperature. It depends on both the properties of the substance and the frequency. Kirchhoff's law states that for any substance, its emissivity at a given wavelength and temperature equals its absorptivity. Note that the absorptivity and emissivity of a given substance may be quite variable for different frequencies.

As seen in Eq. (5.1), the total radiation from a blackbody is dependent on the fourth power of its absolute temperature. The frequency of the maximum intensity of this radiation is also related to temperature through Wien's displacement law (derived from Planck's law):

$$v_{\mathrm{max}} = 1.04 \times 10^{11}\, T \tag{5.2}$$

where frequency v is in s^{-1} and the constant is in s^{-1} K^{-1}.

The radiant flux can be determined as a function of frequency from Planck's distribution law for emission:

$$E_v \mathrm{d}v = c_1 v^3 \left[\exp(c_2 v/T) - 1\right]^{-1} \mathrm{d}v \tag{5.3}$$

where

$$\begin{aligned} c_1 &= 2\pi h/c^2 \\ h &= 6.55 \times 10^{-27} \text{ erg s (Planck's constant)} \\ c &= 3 \times 10^8 \text{ m s}^{-1} \text{ (speed of light)} \\ c_2 &= h/k \end{aligned}$$

and

$$k = 1.37 \times 10^{-16} \text{ erg K}^{-1} \text{ (Boltzmann's constant)}$$

The radiation from a blackbody is continuous over the electromagnetic spectrum. The use of the term black in blackbody, which implies a particular color, is quite misleading, as a number of nonblack materials approach blackbodies in behavior. The sun behaves almost like a blackbody; snow radiates in the infrared nearly as a blackbody. At some wavelengths, water vapor radiates very efficiently. Unlike solids and liquids, many gases absorb (and reradiate) selectively in discrete wavelength bands, rather than smoothly over a continuous spectrum.

B. Incoming Solar Radiation

The sun radiates approximately as a blackbody, with an effective temperature of about 6000 K. The total solar flux is 3.9×10^{26} W. Using Wien's law, it has been found that the frequency of maximum solar radiation intensity is $6.3 \times 10^{14}\ s^{-1}$ ($\lambda = 0.48\,\mu m$), which is in the visible part of the spectrum; 99% of solar radiation occurs between the frequencies of $7.5 \times 10^{13}\ s^{-1}$ ($\lambda = 4\,\mu m$) and $2 \times 10^{15}\ s^{-1}$ ($\lambda = 0.15\,\mu m$) and about 50% in the visible region between $4.3 \times 10^{14}\ s^{-1}$ ($\lambda = 0.7\,\mu m$) and $7.5 \times 10^{14}\ s^{-1}$ ($\lambda = 0.4\,\mu m$). The intensity of this energy flux at the distance of the earth is about 1400 W m^{-2} on an area normal to a beam of solar radiation. This value is called the *solar constant*. Due to the eccentricity of the earth's orbit as it revolves around the sun once a year, the earth is closer to the sun in January (perihelion) than in July (aphelion). This results in about a 7% difference in radiant flux at the outer limits of the atmosphere between these two times.

Since the area of the solar beam intercepted by the earth is πE^2, where E is the radius of the earth, and the energy falling within this circle is spread over the area of the earth's sphere, $4\pi E^2$, in 24 h, the average energy reaching the top of the atmosphere is 338 W m^{-2}. This average radiant energy reaching the outer limits of the atmosphere is depleted as it attempts to reach the earth's surface. Ultraviolet radiation with a wavelength less than 0.18 μm is strongly absorbed by molecular oxygen in the ionosphere 100 km above the earth; shorter X-rays are absorbed at even higher altitudes above the earth's surface. At 60–80 km above the earth, the absorption of 0.2–0.24 μm wavelength radiation leads to the formation of ozone; below 60 km there is so much ozone that much of the 0.2–0.3 μm wavelength radiation is absorbed. This ozone layer in the lower mesosphere and the top of the stratosphere shields life from much of the harmful ultraviolet radiation. The various layers warmed by the absorbed radiation reradiate in wavelengths dependent on their temperature and spectral emissivity. Approximately 5% of the total incoming solar radiation is absorbed above 40 km. Under clear sky conditions, another 10–15% is absorbed by the lower atmosphere or scattered back to space by the atmospheric aerosols and molecules; as a result, only 80–85% of the incoming radiation reaches the earth's surface. With average cloudiness, only about 50% of the incoming radiation reaches the earth's surface, because of the additional interference of the clouds.

C. Albedo and Angle of Incidence

The portion of the incoming radiation reflected and scattered back to space is the *albedo*. The albedo of clouds, snow, and ice-covered surfaces is around

TABLE 5.1

Percent of Incident Radiation Reflected by a Water Surface (Albedo of Water)[a]

Angle of incidence	Percent reflected	Percent absorbed
90	2.0	98.0
70	2.1	97.9
50	2.5	97.5
40	3.4	96.6
30	6.0	94.0
20	13.0	87.0
10	35.0	65.0
5	58.0	42.0

[a] Adapted from Fig. 3.13 of Battan [1].

0.5–0.8, that of fields and forests is 0.03–0.3, and that of water is 0.02–0.05 except when the angle of incidence becomes nearly parallel to the water surface. Table 5.1 shows the albedo of a water surface as a function of the angle of incidence. The albedo averaged over the earth's surface is about 0.35.

Although events taking place on the sun, such as sunspots and solar flares, alter the amount of radiation, the alteration is almost entirely in the X-ray and ultraviolet regions and does not affect the amount in the wavelengths reaching the earth's surface. Therefore, the amount of radiation from the sun that can penetrate to the earth's surface is remarkably constant.

In addition to the effect of albedo on the amount of radiation that reaches the earth's surface, the angle of incidence of the radiation compared to the perpendicular to the surface affects the amount of radiation flux on an area. The flux on a horizontal surface S_h is as follows:

$$S_\mathrm{h} = S \cos Z \tag{5.4}$$

where S is the flux through an area normal to the solar beam and Z is the zenith angle (between the local vertical, the zenith, and the solar beam).

Because of the tilt of the earth's axis by 23.5° with respect to the plane of the earth's revolution around the sun, the north pole is tilted toward the sun on June 22 and away from the sun on December 21 (Fig. 5.2). This tilt causes the solar beam to have perpendicular incidence at different latitudes depending on the date. The zenith angle Z is determined from:

$$\cos Z = \sin \phi \sin \delta + \cos \phi \cos \delta \cos \eta \tag{5.5}$$

where ϕ is latitude (positive for Northern Hemisphere, negative for Southern Hemisphere), δ is solar declination (see Table 5.2), and η is hour angle, 15° × the number of hours before or after local noon.

The solar azimuth ω is the angle between south and the direction toward the sun in a horizontal plane:

$$\sin \omega = (\cos \delta \sin \eta)/\sin Z \tag{5.6}$$

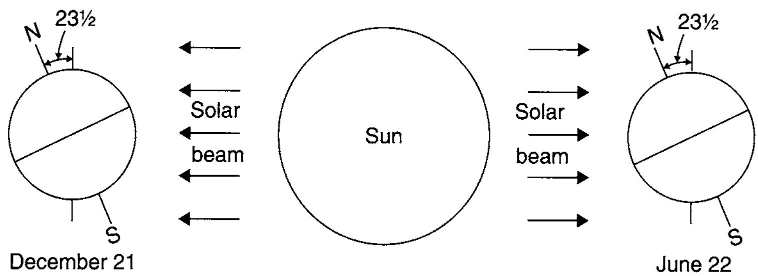

Fig. 5.2. Orientation of the earth to the solar beam at the extremes of its revolution around the sun.

TABLE 5.2

Solar Declination[a]

Date	Declination degree	Date	Declination degree
January 21	−20.90	July 21	20.50
February 21	−10.83	August 21	12.38
March 21	0.0	September 21	1.02
April 21	11.58	October 21	−10.42
May 21	20.03	November 21	−19.75
June 21	23.45	December 21	−23.43

[a] Adapted from Table 2.1 of Byers [2].

Since many surfaces receiving sunlight are not horizontal, a slope at an angle i from the horizontal facing an azimuth ω' degrees from south experiences an intensity of sunlight (neglecting the effects of the atmosphere) of

$$S_s = S[\cos Z \cos i + \sin Z \sin i \cos(\omega - \omega')] \tag{5.7}$$

Here ω and ω' are negative to the east of south and positive to the west.

At angles away from the zenith, solar radiation must penetrate a greater thickness of the atmosphere. Consequently, it can encounter more scattering due to the presence of particles and greater absorption due to this greater thickness.

D. Outgoing Longwave Radiation

Because most ultraviolet radiation is absorbed from the solar spectrum and does not reach the earth's surface, the peak of the solar radiation which reaches the earth's surface is in the visible part of the spectrum. The earth reradiates nearly as a blackbody at a mean temperature of 290 K. The resulting infrared radiation extends over wavelengths of 3–80 μm, with a peak at around 11 μm. The atmosphere absorbs and reemits this longwave radiation primarily because of water vapor but also because of carbon dioxide in the atmosphere. Because of the absorption spectrum of these gases, the atmosphere is mostly opaque to wavelengths less than 7 μm and greater than 14 μm and partly opaque between 7 and 8.5 μm and between 11 and 14 μm. The atmosphere loses heat to space directly through the nearly transparent window between 8.5 and 11 μm and also through the absorption and successive reradiation by layers of the atmosphere containing these absorbing gases.

Different areas of the earth's surface react quite differently to heating by the sun. For example, although a sandy surface reaches fairly high temperatures on a sunny day, the heat capacity and conductivity of sand are relatively low; the heat does not penetrate more than about 0.2–0.3 m and little heat is stored. In contrast, in a body of water, the sun's rays penetrate several meters and slowly heat a fairly deep layer. In addition, the water can move readily and convection can spread the heat through a deeper layer. The heat capacity of water is considerably greater than that of sand. All these factors combine to allow considerable storage of heat in water bodies.

E. Heat Balance

Because of the solar beam's more direct angle of incidence in equatorial regions, considerably more radiation penetrates and is stored by water near the equator than water nearer the poles. This excess is not compensated for by the outgoing longwave radiation, yet there is no continual buildup of heat in equatorial regions. The first law of thermodynamics requires that the energy entering the system (earth's atmosphere) be balanced with that exiting the system. Figure 5.3 shows the annual mean incoming and outgoing radiation averaged over latitude bands. There is a transfer of heat poleward from the equatorial regions to make up for a net outward transfer of heat

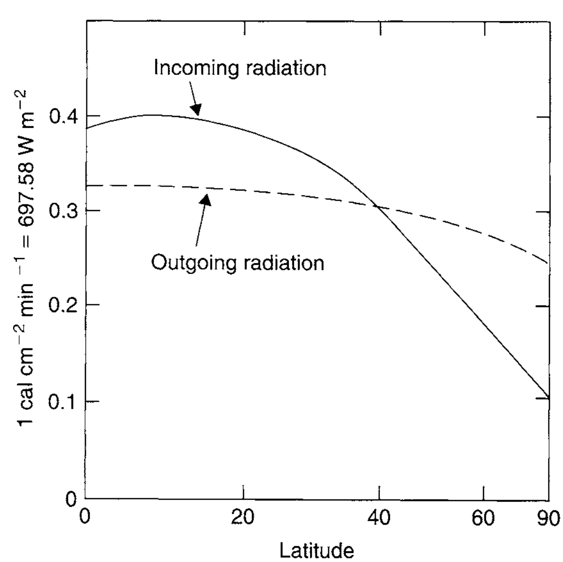

Fig. 5.3. Annual mean radiation by latitude. Note that the latitude scale simulates the amount of the earth's surface area between the latitude bands. Incoming radiation is that absorbed by earth and atmosphere. Outgoing radiation is that leaving the atmosphere. *Source*: After Byers [2] (Note: 1 cal cm^{-2} min^{-1} = 697.58 W m^{-2}).

near the poles. This heat is transferred by air and ocean currents as warm currents move poleward and cool currents move equatorward. Considerable heat transfer occurs by the evaporation of water in the tropics and its condensation into droplets farther poleward, with the release of the heat of condensation. Enough heat is transferred to result in no net heating of the equatorial regions or cooling of the poles. The poleward flux of heat across various latitudes is shown in Table 5.3.

Taking the earth as a whole over a year or longer, because there is no appreciable heating or cooling, there is a heat balance between the incoming solar radiation and the radiation escaping to space. This balance is depicted as bands of frequency of electromagnetic radiation in Fig. 5.4.

TABLE 5.3

Poleward Flux of Heat Across Latitudes (10^{19} kcal year^{-1})[a]

Latitude	Flux	Latitude	Flux
10	1.21	50	3.40
20	2.54	60	2.40
30	3.56	70	1.25
40	3.91	80	0.35

[a] Adapted from Table 12 of Sellers [3].

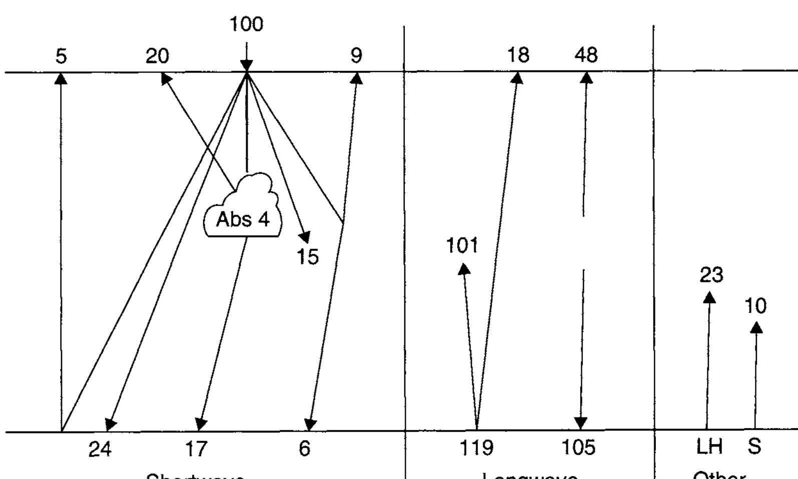

Fig. 5.4. Radiation heat balance. The 100 units of incoming shortwave radiation are distributed: reflected from earth's surface to space, 5; reflected from cloud surfaces to space, 20; direct reaching earth, 24; absorbed in clouds, 4; diffuse reaching earth through clouds, 17; absorbed in atmosphere, 15; scattered to space, 9; scattered to earth, 6. The longwave radiation comes from (1) the earth radiating 119 units: 101 to the atmosphere and 18 directly to space, and (2) the atmosphere radiating 105 units back to earth and 48 to space. Additional transfers from the earth's surface to the atmosphere consist of latent heat (LH), 23; and sensible (S) heat, 10. *Source*: After Lowry [4].

II. MOTION

Vertical air motions affect both weather and the mixing processes of importance to air pollution. Upward vertical motions can be caused by lifting over terrain, lifting over weather fronts, and convergence toward low-pressure centers. Downward vertical motions can be caused by sinking to make up for divergence near high-pressure centers. One must know whether the atmosphere enhances or suppresses these vertical motions to assess their effects. When the atmosphere resists vertical motions, it is called *stable*; when the atmosphere enhances vertical motions, it is called *unstable* or in a state of *instability*.

In incompressible fluids, such as water, the vertical structure of temperature very simply reveals the stability of the fluid. When the lower layer is warmer and thus less dense than the upper layer, the fluid is unstable and convective currents will cause it to overturn. When the lower layer is cooler than the upper layer, the fluid is stable and vertical exchange is minimal. However, because air is compressible, the determination of stability is somewhat more complicated. The temperature and density of the atmosphere normally decrease with elevation; density is also affected by moisture in the air.

The relationship between pressure p, volume V, mass m, and temperature T is given by the equation of state:

$$pV = RmT \tag{5.8}$$

where R is a specific gas constant equal to the universal gas constant divided by the gram molecular weight of the gas. Since the density ρ is m/V, the equation can be rewritten as

$$p = R\rho T \tag{5.9}$$

or considering specific volume $\alpha = 1/\rho$ as

$$\alpha p = RT \tag{5.10}$$

These equations combine Boyle's law, which states that when temperature is held constant the volume varies inversely with the pressure, and the law of Guy-Lussac, which states that when pressure is held constant the volume varies in proportion to the absolute temperature.

A. First Law of Thermodynamics

If a volume of air is held constant and a small amount of heat Δh is added, the temperature of the air will increase by a small amount ΔT. This can be expressed as

$$\Delta h = c_v \, \Delta T \tag{5.11}$$

where c_v is the specific heat at constant volume. In this case, all the heat added is used to increase the internal energy of the volume affected by the temperature. From the equation of state (Eq. 5.8), it can be seen that the pressure will increase.

If, instead of being restricted, the volume of air considered is allowed to remain at an equilibrium constant pressure and expand in volume, as well as change temperature in response to the addition of heat, this can be expressed as

$$\Delta h = c_v \, \Delta T + p \, \Delta v \tag{5.12}$$

By using the equation of state, the volume change can be replaced by a corresponding pressure change:

$$\Delta h = c_p \Delta T + v \, \Delta p \tag{5.13}$$

where c_p is the specific heat at constant pressure and equals $c_v + R_d$, where R_d is the gas constant for dry air.

B. Adiabatic Processes

An adiabatic process is one with no loss or gain of heat to a volume of air. If heat is supplied or withdrawn, the process is *diabatic* or *nonadiabatic*. Near the earth's surface, where heat is exchanged between the earth and the air, the processes are diabatic.

However, away from the surface, processes frequently are adiabatic. For example, if a volume (parcel) of air is forced upward over a ridge, the upward-moving air will encounter decreased atmospheric pressure and will expand and cool. If the air is not saturated with water vapor, the process is called *dry adiabatic*. Since no heat is added or subtracted, Δh in Eq. (5.13) can be set equal to zero, and introducing the hydrostatic equation

$$-\Delta p = \rho g \, \Delta z \tag{5.14}$$

and combining equations results in

$$-\Delta T / \Delta z = g / c_p \tag{5.15}$$

Thus air cools as it rises and warms as it descends. Since we have assumed an adiabatic process, $-\Delta T/\Delta z$ defines γ_d, the dry adiabatic process lapse rate, a constant equal to 0.0098 K/m, is nearly 1 K/100 m or 5.4°F/1000 ft.

If an ascending air parcel reaches saturation, the addition of latent heat from condensing moisture will partially overcome the cooling due to expansion. Therefore, the saturated adiabatic lapse rate (of cooling) γ_w is smaller than γ_d.

C. Determining Stability

By comparing the density changes undergone by a rising or descending parcel of air with the density of the surrounding environment, the enhancement or suppression of the vertical motion can be determined. Since pressure decreases with height, there is an upward-directed pressure gradient force. The force of gravity is downward. The difference between these two forces is the buoyancy force. Using Newton's second law of motion, which indicates that a net force equals an acceleration, the acceleration a of an air parcel at a particular position is given by

$$a = g(T_p - T_e)/T_p \tag{5.16}$$

where g is the acceleration due to gravity ($9.8\,\mathrm{m\,s^{-2}}$), T_p is the temperature of an air parcel that has undergone a temperature change according to the process lapse rate, and T_e is the temperature of the surrounding environment at the same height. (Temperatures are expressed in degrees Kelvin.)

Figure 5.5 shows the temperature change undergone by a parcel of air forced to rise 200 m in ascending a ridge. Assuming that the air is dry, and therefore that no condensation occurred, this figure also represents the warming of the air parcel if the flow is reversed so that the parcel moves downslope from B to A.

Comparing the temperature of this parcel to that of the surrounding environment (Fig. 5.6), it is seen that in rising from 100 to 300 m, the parcel

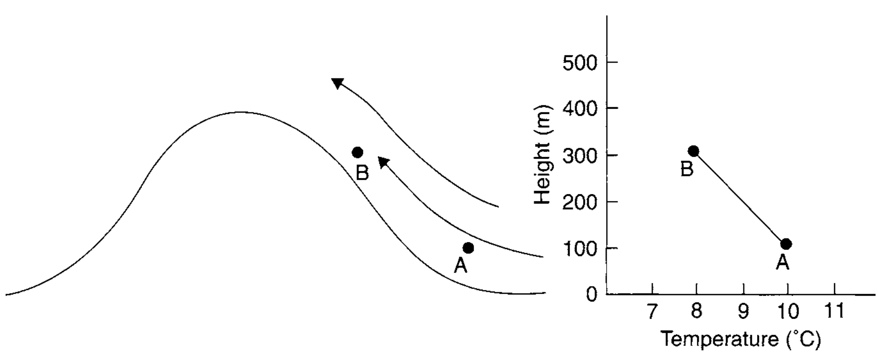

Fig. 5.5. Cooling of ascending air. Dry air forced to rise 200 m over a ridge cools adiabatically by 2°C.

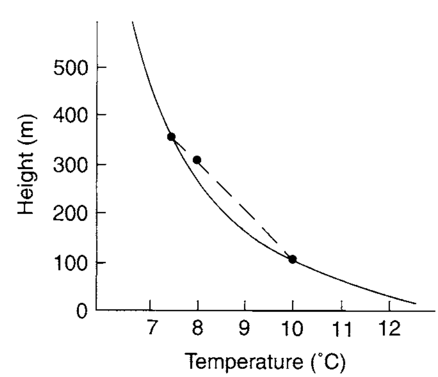

Fig. 5.6. Temperature of a parcel of air forced to rise 200 m compared to the superadiabatic environmental lapse rate. Since the parcel is still warmer than the environment, it will continue to rise.

undergoes the temperature change of the dry adiabatic process lapse rate. The dashed line is a dry adiabatic line or dry adiabat. Suppose that the environmental temperature structure is shown by the solid curve. Since the lapse rate of the surrounding environment in the lowest 150–200 m is steeper than the adiabatic lapse rate (superadiabatic)—that is, since the temperature drops more rapidly with height—this part of the environment is thermally unstable. At 300 m the parcel is 0.2°C warmer than the environment, the resulting acceleration is upward, and the atmosphere is enhancing the vertical motion and is unstable. The parcel of air continues to rise until it reaches 350 m, where its temperature is the same as that of the environment and its acceleration drops to zero. However, above 350 m the lapse rate of the surrounding environment is not as steep as the adiabatic lapse rate (subadiabatic), and this part of the environment is thermally stable (it resists upward or downward motion).

If the temperature structure, instead of being that of Fig. 5.6, differs primarily in the lower layers, it resembles Fig. 5.7, where a temperature inversion (an increase rather than a decrease of temperature with height) exists. In the forced ascent of the air parcel up the slope, dry adiabatic cooling produces parcel temperatures that are everywhere cooler than the environment; acceleration is downward, resisting displacement; and the atmosphere is stable.

Thermodynamic diagrams which show the relationships between atmospheric pressure (rather than altitude), temperature, dry adiabatic lapse rates, and moist adiabatic lapse rates are useful for numerous atmospheric thermodynamic estimations. The student is referred to a standard text on meteorology (see Suggested Reading) for details. In air pollution meteorology, the thermodynamic diagram may be used to determine the current mixing height (the top of the neutral or unstable layer). The mixing height at a given time may be estimated by use of the morning radiosonde ascent plotted on a thermodynamic chart. The surface temperature at the given time is plotted on the diagram. If a dry adiabat is drawn through this temperature, the height

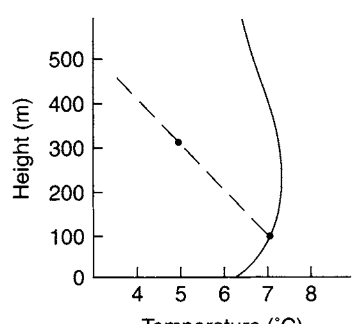

Fig. 5.7. Temperature of a parcel of air forced to rise 200 m compared to an inversion environmental lapse rate. Since the parcel is cooler than the environment, it will sink back to its original level.

aboveground at the point where this dry adiabat intersects the morning sounding is the mixing height for that time. The mixing height for the time of maximum temperature is the maximum mixing height. Use of this sounding procedure provides an approximation because it assumes that there has been no significant advection since the time of the sounding.

D. Potential Temperature

A useful concept in determining stability in the atmosphere is *potential temperature.* This is a means of identifying the dry adiabat to which a particular atmospheric combination of temperature and pressure is related. The potential temperature θ is found from:

$$\theta = T(1000/p)^{0.288} \tag{5.17}$$

where T is temperature and p is pressure (in millibars, mb). This value is the same as the temperature that a parcel of dry air would have if brought dry adiabatically to a pressure of 1000 mb.

If the potential temperature decreases with height, the atmosphere is unstable. If the potential temperature increases with height, the atmosphere is stable. The average lapse rate of the atmosphere is about 6.5°C km^{-1}; that is, the potential temperature increases with height and the average state of the atmosphere is stable.

E. Effect of Mixing

The mixing of air in a vertical layer produces constant potential temperature throughout the layer. Such mixing is usually mechanical, such as air movement over a rough surface. In Fig. 5.8 the initial temperature structure is subadiabatic (solid line). The effect of mixing is to achieve a mean potential temperature throughout the layer (dashed line), which in the lower part

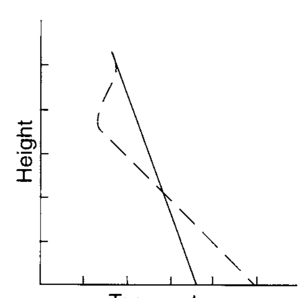

Fig. 5.8. Effect of forced mixing (dashed line) on the environmental subadiabatic lapse rate (solid line). Note the formation of an inversion at the top of the mixed layer.

is dry adiabatic. The bottom part of the layer is warmed; the top is cooled. Note that above the vertical extent of the mixing, an inversion is formed connecting the new cooled portion with the old temperature structure above the zone of mixing. If the initial layer has considerable moisture, although not saturated, cooling in the top portion of the layer may decrease the temperature to the point where some of the moisture condenses, forming clouds at the top. An example of this is the formation of an inversion and a layer of stratus clouds along the California coast.

F. Radiation or Nocturnal Inversions

An inversion caused by mixing in a surface layer was just discussed above. Inversions at the surface are caused frequently at night by radiational cooling of the ground, which in turn cools the air near it.

G. Subsidence Inversions

There is usually some descent (subsidence) of air above surface high-pressure systems. This air warms dry adiabatically as it descends, decreasing the relative humidity and dissipating any clouds in the layer. A subsidence inversion forms as a result of this sinking. Since the descending air compresses as it encounters the increased pressures lower in the atmosphere, the top portion of the descending layer will be further warmed due to its greater descent than will the bottom portion of the layer (Fig. 5.9). Occasionally a subsidence inversion descends all the way to the surface, but usually its base is well above the ground.

Inversions are of considerable interest in relation to air pollution because of their stabilizing influence on the atmosphere, which suppresses the vertical motion that causes the vertical spreading of pollutants.

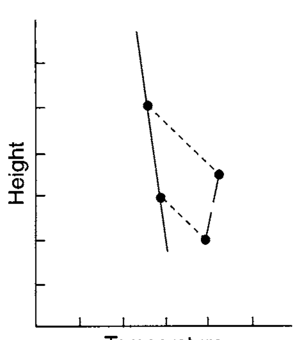

Fig. 5.9. Formation of a subsidence inversion in subsiding (sinking) air. Note the vertical compression of the sinking layer which is usually accompanied by horizontal divergence.

III. ENERGY-MOTION RELATIONSHIPS

The atmosphere is nearly always in motion. The scales and magnitude of these motions extend over a wide range. Although vertical motions certainly occur in the atmosphere and are important to both weather processes and the movement of pollutants, it is convenient to consider wind as only the horizontal component of velocity.

On the regional scale (hundreds to thousands of kilometers), the winds are most easily understood by considering the balance of various forces in the atmosphere. The applicable physical law is Newton's second law of motion, $F = ma$; if a force F is exerted on a mass m, the resulting acceleration a equals the force divided by the mass. This can also be stated as the rate of change of momentum of a body, which is equal to the sum of the forces that act on the body. It should be noted that all the forces to be discussed are vectors; that is, they have both magnitude and direction. Although Newton's second law applies to absolute motion, it is most convenient to consider wind relative to the earth's surface. These create some slight difficulties, but they can be rather easily managed.

A. Pressure Gradient Force

Three forces of importance to horizontal motion are the pressure gradient force, gravity, and friction. Atmospheric pressure equals mass times the acceleration of gravity. Considering a unit volume, $p = \rho g$; the gravitational force on the unit volume is directed downward. Primarily because of horizontal temperature gradients, there are horizontal density gradients and consequently horizontal pressure gradients. The horizontal pressure gradient force $p_h = \Delta p/\rho\Delta x$, where Δp is the horizontal pressure difference over the distance Δx. The direction of this force and of the pressure difference measurement is locally perpendicular to the lines of equal pressure (isobars) and is directed from high to low pressure.

B. Coriolis Force

If the earth were not rotating, the wind would blow exclusively from high to low pressure. Close to the earth, it would be slowed by friction between the atmosphere and the earth's surface but would maintain the same direction with height. However, since the earth undergoes rotation, there is an apparent force acting on horizontal atmospheric motions when examined from a point of reference on the earth's surface. For example, consider a wind of velocity $10\,\mathrm{m\,s^{-1}}$ blowing at time 1 in the direction of the 0° longitude meridian across the north pole (Fig. 5.10). The wind in an absolute sense continues to blow in this direction for 1 h, and a parcel of air starting at the pole at time 1 travels 36 km in this period. However, since the earth turns 360° every 24 h, or $15°\,\mathrm{h^{-1}}$, it has rotated 15° in the hour and we find that at time 2 (60 min

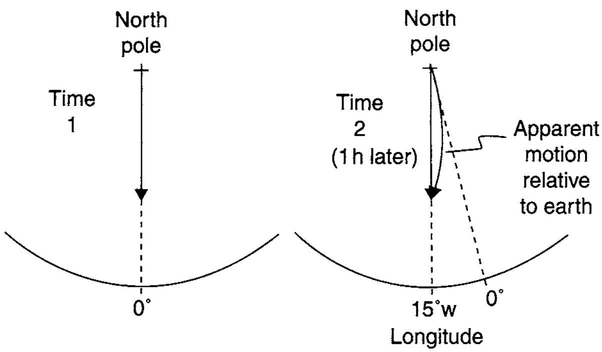

Fig. 5.10. Effect of the coriolis force. The path of air moving from the north pole to the south as viewed from space is straight; as viewed from the earth's surface it is curved.

after time 1) the 15° meridian is now beneath the wind vector. As viewed from space (the absolute frame of reference), the flow has continued in a straight line. However, as viewed from the earth, the flow has undergone an apparent deflection to the right. The force required to produce this apparent deflection is the coriolis force and is equal to $D = vf$ where f, the coriolis parameter, equals $2\Omega \sin \phi$. Here Ω is the angular speed of the earth's rotation, $2\pi/(24 \times 60 \times 60) = 7.27 \times 10^{-5}\,\text{s}^{-1}$, and ϕ is the latitude. It is seen that f is maximal at the poles and zero at the equator. The deflecting force is to the right of the wind vector in the Northern Hemisphere and to the left in the Southern Hemisphere. For the present example, the deflecting force is $1.45 \times 10^{-3}\,\text{ms}^{-2}$, and the amount of deflection after the 36-km movement in 1 h is 9.43 km.

C. Geostrophic Wind

Friction between the atmosphere and the earth's surface may generally be neglected at altitudes of about 700 m and higher. Therefore, large-scale air currents closely represent a balance between the pressure gradient force and the coriolis force. Since the coriolis force is at a right angle to the wind vector, when the coriolis force is equal in magnitude and opposite in direction to the pressure gradient force, a wind vector perpendicular to both of these forces occurs, with its direction along the lines of constant pressure (Fig. 5.11). In the Northern Hemisphere, the low pressure is to the left of the wind vector (Buys Ballot's law); in the Southern Hemisphere, low pressure is to the right. The geostrophic velocity is

$$v_g = -\Delta p/\rho f\,\Delta d \tag{5.18}$$

When the isobars are essentially straight, the balance between the pressure gradient force and the coriolis force results in a geostrophic wind parallel to the isobars.

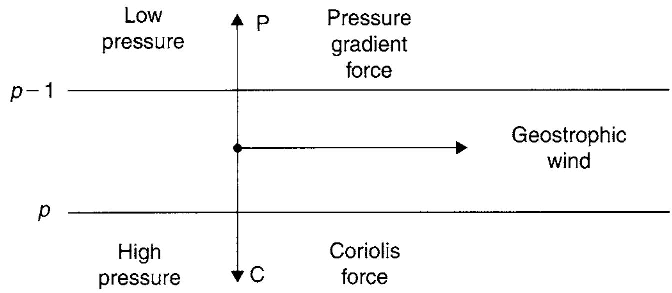

Fig. 5.11. Balance of forces resulting in geostrophic wind.

D. Gradient Wind

When the isobars are curved, an additional force, a centrifugal force outward from the center of curvature, enters into the balance of forces. In the case of curvature around low pressure, a balance of forces occurs when the pressure gradient force equals the sum of the coriolis and centrifugal forces (Fig. 5.12) and the wind continues parallel to the isobars. In the case of curvature around high pressure, a balance of forces occurs when the sum of the pressure gradient and centrifugal forces equals the coriolis force (Fig. 5.13). To maintain a given gradient wind speed, a greater pressure gradient force (tighter spacing of the isobars) is required in the flow around low-pressure systems than in the flow around high-pressure systems.

E. The Effect of Friction

The frictional effect of the earth's surface on the atmosphere increases as the earth's surface is approached from aloft. Assuming that we start with geostrophic balance aloft, consider what happens to the wind as we move downward toward the earth. The effect of friction is to slow the wind velocity, which in turn decreases the coriolis force. The wind then turns toward low pressure until the resultant vector of the frictional force and the coriolis

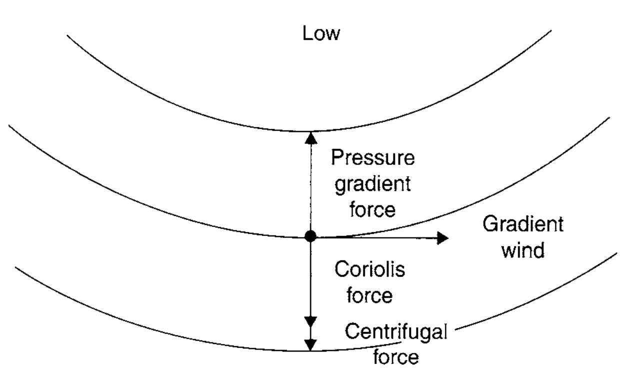

Fig. 5.12. Balance of forces resulting in gradient wind around low pressure.

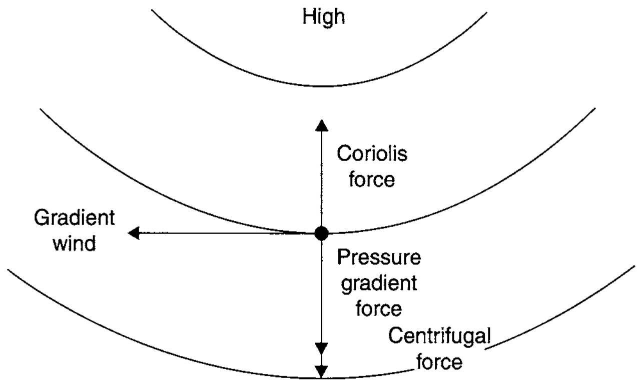

Fig. 5.13. Balance of forces resulting in gradient wind around high pressure. Note that the wind speed is greater for a given pressure gradient force than that around low pressure.

force balances the pressure gradient force (Fig. 5.14). The greater the friction, the slower the wind and the greater the amount of turning toward low pressure. The turning of the wind from the surface through the friction layer is called the *Ekman spiral*. A radial plot, or hodograph, of the winds through the friction layer is shown diagrammatically in Fig. 5.15.

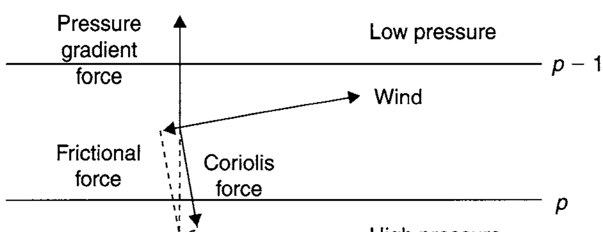

Fig. 5.14. Effect of friction on the balance of forces, causing wind to blow toward low pressure.

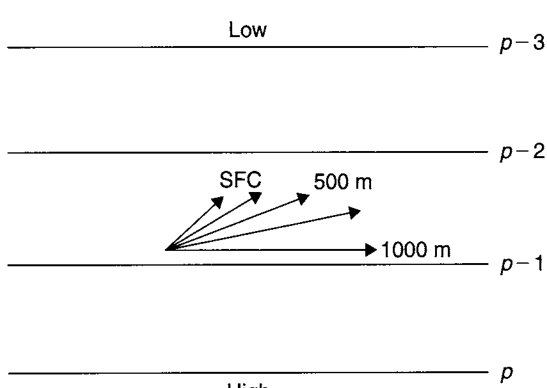

Fig. 5.15. Hodograph showing variation of wind speed and direction with height above ground. SFC: surface wind.

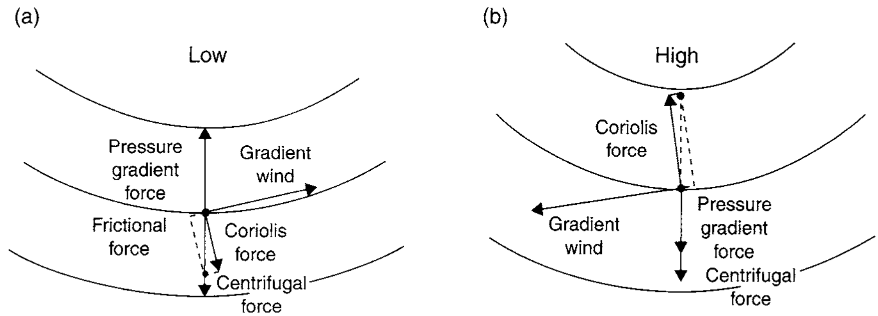

Fig. 5.16. Effect of friction upon gradient wind around (a) low and (b) high pressures.

Note that this frictional effect will cause pollutants released at two different heights to tend to move in different directions.

In the friction layer where the isobars are curved, the effect of frictional drag is added to the forces discussed under gradient wind. The balance of the pressure gradient force, the coriolis deviating force, the centrifugal force, and the frictional drag in the vicinity of the curved isobars results in wind flow around low pressure and high pressure in the Northern Hemisphere, as shown in Fig. 5.16.

F. Vertical Motion: Divergence

So far in discussing motion in the atmosphere, we have been emphasizing only horizontal motions. Although of much smaller magnitude than horizontal motions, vertical motions are important both to daily weather formation and to the transport and dispersion of pollutants.

Persistent vertical motions are linked to the horizontal motions. If there is divergence (spreading) of the horizontal flow, there is sinking (downward vertical motion) of air from above to compensate. Similarly, converging (negative divergence) horizontal air streams cause upward vertical motions, producing condensation and perhaps precipitation in most air masses, as well as transport of air and its pollutants from near the surface to higher altitudes.

IV. LOCAL WIND SYSTEMS

Frequently, local wind systems are superimposed on the larger-scale wind systems just discussed. These local flows are especially important to air pollution since they determine the amount of a pollutant that will come in contact with the receptor. In fact, local conditions may dominate when the larger-scale flow becomes light and indefinite. Local wind systems are usually quite significant in terms of the transport and dispersion of air pollutants.

A. Sea and Land Breezes

The sea breeze is a result of the differential heating of land and water surfaces by incoming solar radiation. Since solar radiation penetrates several meters of a body of water, it warms very slowly. In contrast, only the upper few centimeters of land are heated, and warming occurs rapidly in response to solar heating. Therefore, especially on clear summer days, the land surface heats rapidly, warming the air near the surface and decreasing its density. This causes the air to rise over the land, decreasing the atmospheric pressure near the surface relative to the pressure at the same altitude over the water surface. The rising air increases the pressure over the land relative to that above the water at altitudes of approximately 100–200 m. The air that rises over the land surface is replaced by cooler air from over the water surface. This air, in turn, is replaced by subsiding air from somewhat higher layers of the atmosphere over the water. Air from the higher-pressure zone several hundred meters above the surface then flows from over the land surface out over the water, completing a circular or cellular flow (Fig. 5.17). Any general flow due to large-scale pressure systems will be superimposed on the sea breeze and may either reinforce or inhibit it. Ignoring the larger-scale influences, the strength of the sea breeze will generally be a function of the temperature excess of the air above the land surface over that above the water surface.

Just as heating in the daytime occurs more quickly over land than over water, at night radiational cooling occurs more quickly over land. The pressure pattern tends to be the reverse of that in the daytime. The warmer air tends to rise over the water, which is replaced by the land breeze from land to water, with the reverse flow (water to land) completing the circular flow at altitudes somewhat aloft. Frequently at night, the temperature differences between land and water are smaller than those during the daytime, and therefore the land breeze has a lower speed.

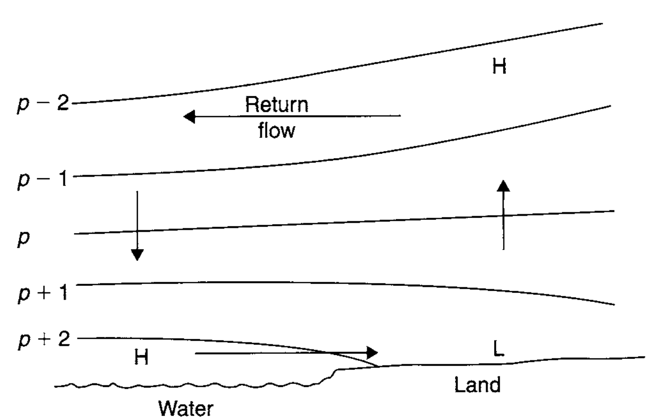

Fig. 5.17. Sea breeze due to surface heating over land, resulting in thermals, and subsidence over water.

B. Mountain and Valley Winds

Solar heating and radiational cooling influence local flows in terrain situations. Consider midday heating of a south-facing mountainside. As the slope heats, the air adjacent to the slope warms, its density is decreased, and the air attempts to ascend (Fig. 5.18). Near the top of the slope, the air tends to rise vertically. Along each portion of the slope farther down the mountain, it is easier for each rising parcel of air to move upslope, replacing the parcel ahead of it rather than rising vertically. This upslope flow is the valley wind.

At night when radiational cooling occurs on slopes, the cool dense air near the surface descends along the slope (Fig. 5.19). This is the downslope wind. To compensate for this descending air, air farther from the slope that is cooled very little is warmer relative to the descending air and rises, frequently resulting in a closed circular path. Where the downslope winds occur on opposite slopes of a valley, the cold air can accumulate on the valley floor. If there is any slope to the valley floor, this pool of cold air can move down the valley, resulting in a drainage or canyon wind.

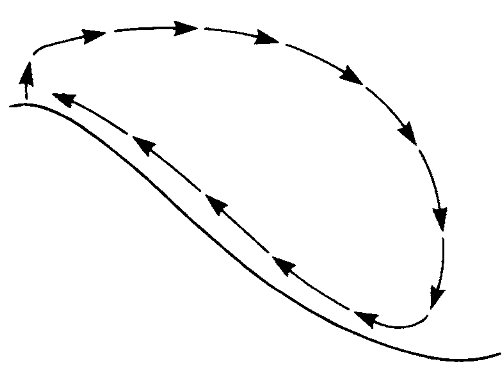

Fig. 5.18. Upslope wind (daytime) due to greater solar heating on the valley's side than in its center.

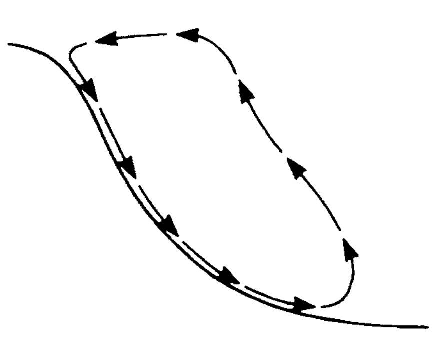

Fig. 5.19. Downslope wind (night) due to more rapid radiational cooling on the valley's slope than in its center.

Different combinations of valley and mountain slope, especially with some slopes nearly perpendicular to the incoming radiation and others in deep shadow, lead to many combinations of wind patterns, many nearly unique. Also, each local flow can be modified by the regional wind at the time which results from the current pressure patterns. Table 5.4 gives characteristics of eight different situations depending on the orientation of the ridgeline and valley with respect to the sun, wind direction perpendicular or parallel to the ridgeline, and time of day. Figure 5.20 shows examples of some of the mountain and valley winds listed in Table 5.4. These are rather idealized circulations compared to observed flows at any one time.

The effect of solar radiation is different with valley orientation. An east–west valley has only one slope that is significantly heated—the south-facing slope may be near normal with midday sunshine. A north–south valley will have both slopes heated at midday. The effect of flow in relation to valley orientation is such that flows perpendicular to valleys tend to form circular eddies and encourage local flows; flows parallel to valleys tend to discourage local flows and to sweep clean the valley, especially with stronger wind speeds.

Keep in mind that the flows occurring result from the combination of the general and local flows; the lighter the general flow, the greater the opportunity for generation of local flows.

TABLE 5.4

Generalized Mesoscale Windflow Patterns Associated with Different Combinations of Wind Direction and Ridgeline Orientation

Wind direction relative to ridgeline	Time of day	Ridgeline orientation: East–west	Ridgeline orientation: North–south
Parallel	Day	1[a] South-facing slope is heated—single helix	2 Upslope flow on both heated slopes—double helix
	Night	3 Downslope flow on both slopes—double helix	4 Downslope flow on both slopes—double helix
Perpendicular	Day	South-facing slope is heated 5a North wind—stationary eddy fills valley 5b South wind—eddy suppressed, flow without separation	6 Upslope flow on both heated slopes—stationary eddy on one side of valley
	Night	7 Indefinite flow—extreme stagnation in valley bottom	8 Indefinite flow—extreme stagnation in valley bottom

[a] Numbers refer to Fig. 5.20.

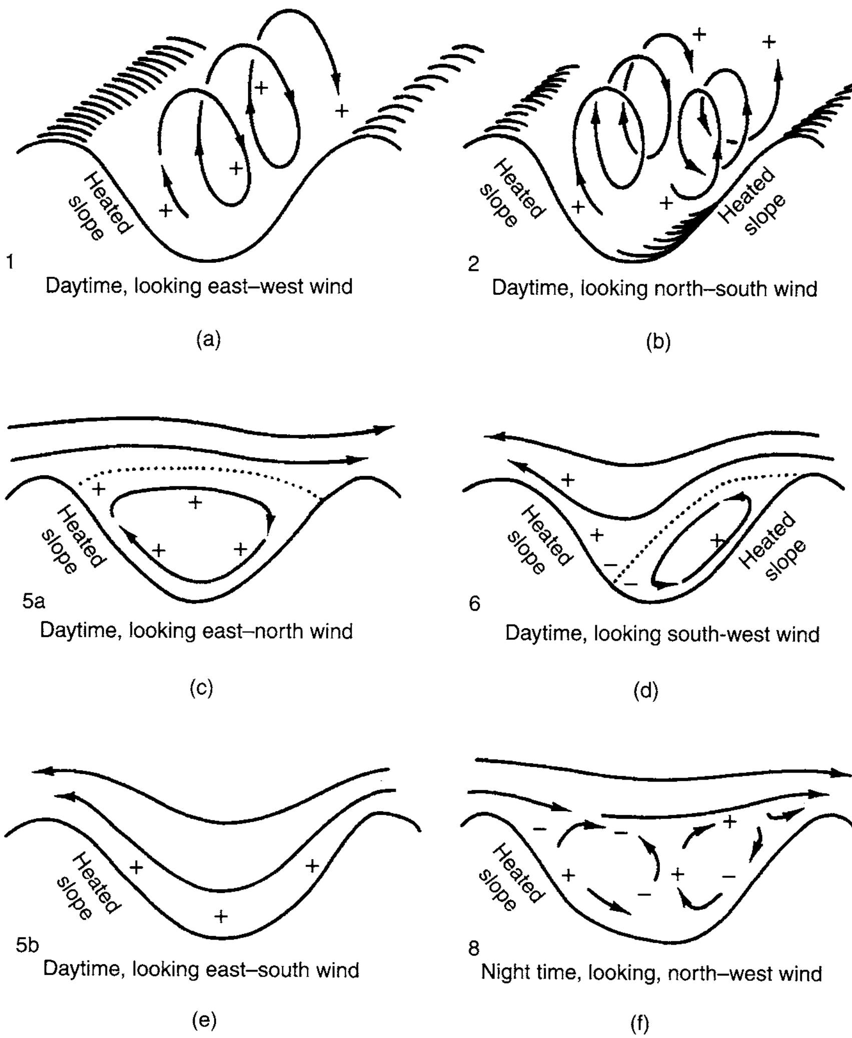

Fig. 5.20. Local valley–ridge flow patterns (numbers refer to Table 5.4).

Complicated terrain such as a major canyon with numerous side canyons will produce complicated and unique flows, especially when side canyon drainage flows reinforce the drainage flow in the main valley.

C. Urban–Rural Circulations

Urban areas have roughness and thermal characteristics different from those of their rural surroundings. Although the increased roughness affects both the vertical wind profile and the vertical temperature profile, the effects due to the thermal features are dominant. The asphalt, concrete, and steel of urban areas heat quickly and have a high heat-storing capability compared to the soil and vegetation of rural areas. Also, some surfaces of buildings are normal to the sun's rays just after sunrise and also before sunset, allowing warming throughout the day. The result is that the urban area becomes warmer than its surroundings during the day and stores sufficient heat that reradiation of the stored heat during the night keeps the urban atmosphere considerably warmer than its rural surroundings throughout most nights with light winds.

Under the lightest winds, the air rises over the warmest part of the urban core, drawing cooler air from all directions from the surroundings (Fig. 5.21). Subsidence replaces this air in rural areas, and a closed torus (doughnut)-shaped circulation occurs with an outflow above the urban area. This circulation is referred to as the *urban heat island*. The strength of the resulting flow is dependent on the difference in temperature between the urban center and its surroundings.

When the regional wind allows the outflow to take place in primarily one direction and the rising warm urban air moves off with this regional flow, the circulation is termed the *urban plume* (Fig. 5.22). Under this circumstance, the inflow to the urban center near the surface may also be asymmetric, although it is more likely to be symmetric than the outflow at higher altitudes.

The urban area also gives off heat through the release of gases from combustion and industrial processes. Compared to the heat received through solar radiation and subsequently released, the combustion and process heat

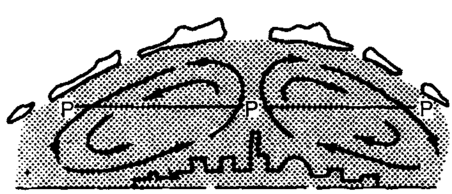

Fig. 5.21. Urban heat island (light regional wind).

Fig. 5.22. Urban plume (moderate regional wind).

is usually quite small, although it may be 10% or more in major urban areas. It can be of significance in the vicinity of a specific local source, such as a steam power plant (where the release of heat is large over a small area) and during light-wind winter conditions.

D. Flow Around Structures

When the wind encounters objects in its path such as an isolated structure, the flow usually is strongly perturbed and a turbulent wake is formed in the vicinity of the structure, especially downwind of it. If the structure is semi-streamlined in shape, the flow may move around it with little disturbance. Since most structures have edges and corners, generation of a turbulent wake is quite common. Figure 5.23 shows schematically the flow in the vicinity of a cubic structure. The disturbed flow consists of a cavity with strong turbulence and mixing, a wake extending downwind from the cavity a distance equivalent to a number of structure side lengths, a displacement zone where flow is initially displaced before entering the wake, and a region of flow that is displaced away from the structure but does not get caught in the wake. Wind tunnels, water channels, and/or towing tanks are extremely useful in studying building wake effects.

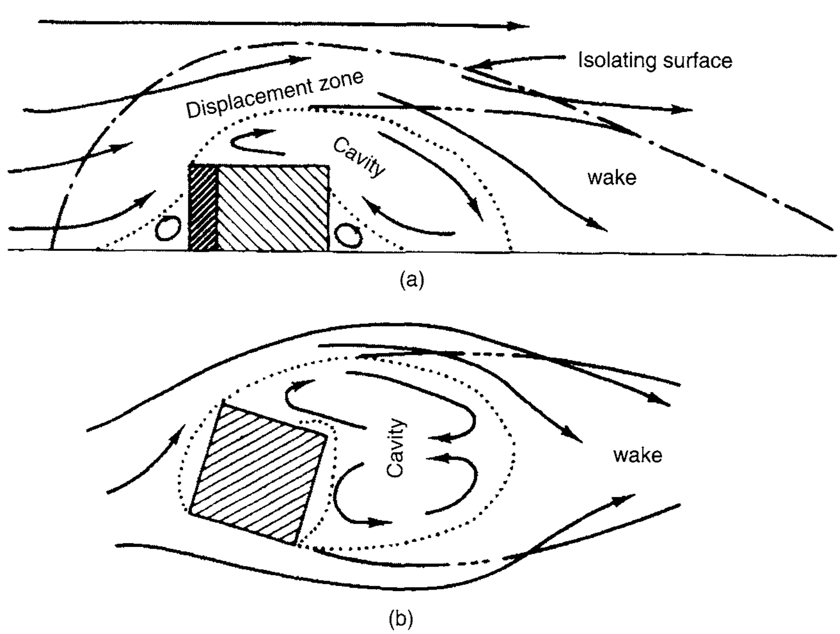

Fig. 5.23. Aerodynamic flow around a cube: (a) side view and (b) plan view. *Source*: After Halitsky [5].

V. GENERAL CIRCULATION

Atmospheric motions are driven by the heat from incoming solar radiation and the redistribution and dissipation of this heat to maintain constant temperatures on the average. The atmosphere is inefficient, because only about 2% of the received incoming solar radiation is converted to kinetic energy, that is, air motion; even this amount of energy is tremendous compared to that which humans are able to produce. As was shown in Section I, a surplus of radiant energy is received in the equatorial regions and a net outflux of energy occurs in the polar regions. Many large-scale motions serve to transport heat poleward or cooler air toward the equator.

If the earth did not rotate or if it rotated much more slowly than it does, a meridional (along meridians) circulation would take place in the troposphere (Fig. 5.24). Air would rise over the tropics, move poleward, sink over the poles forming a subsidence inversion, and then stream equatorward near the earth's surface. However, since the earth's rotation causes the apparent deflection due to the coriolis force, meridional motions are deflected to become zonal (along latitude bands) before moving more than 30°. Therefore, instead of the single cell consisting of dominantly meridional motion (Fig. 5.24), meridional transport is accomplished by three cells between the equator and the pole (Fig. 5.25). This circulation results in subsidence inversions and high pressure where there is sinking toward the earth's surface and low pressure where there is upward motion.

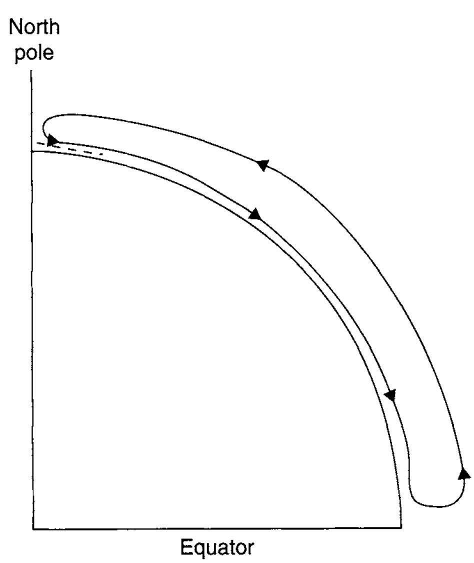

Fig. 5.24. Meridional single-cell circulation (on the sunny side of a nonrotating earth).

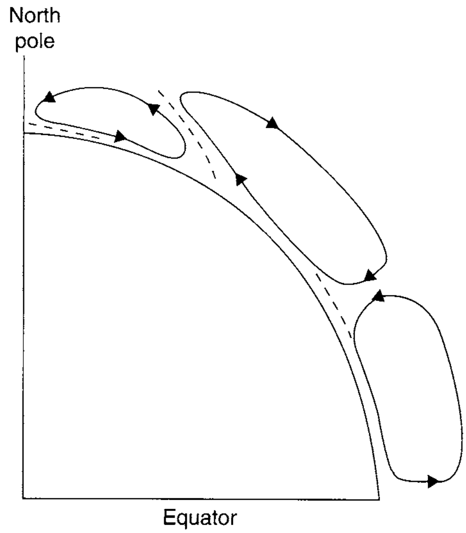

Fig. 5.25. Meridional three-cell circulation (rotating earth).

A. Tropics

Associated with the cell nearest the equator are surface winds moving toward the equator which are deflected toward the west. In the standard terminology of winds, which uses the direction from which they come, these near-surface winds are referred to as *easterlies* (Fig. 5.26), also called *trade*

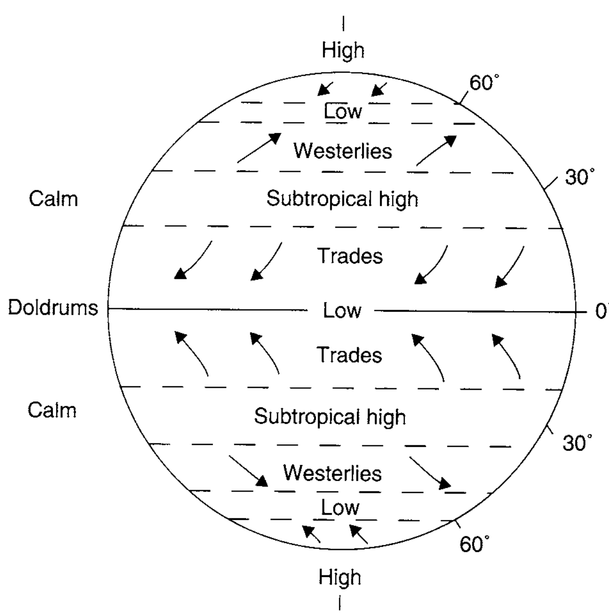

Fig. 5.26. Near-surface winds for various latitude belts.

winds. Since most of the earth's surface near the equator is ocean, these winds absorb heat and moisture on their way to the equator.

Where the trade winds from each hemisphere meet is a low-pressure zone, the *intertropical convergence zone*. This zone of light winds or doldrums shifts position with the season, moving slightly poleward into the summer hemisphere. The rising air with high humidity in the convective motions of the convergence zone causes heavy rainfall in the tropics. This giant convective cell, or Hadley cell, absorbs heat and the latent heat of evaporation at low levels, releasing the latent heat as the moisture condenses in the ascending air. Some of this heat is lost through infrared radiation from cloud tops. The subsiding air, which warms adiabatically as it descends in the vicinity of 30° latitude (horse latitudes), feeds warm air into the mid-latitudes. Although the position of the convergence zone shifts somewhat seasonally, the Hadley cell circulation is quite persistent, resulting in a fairly steady circulation.

B. Mid-latitudes

Because at higher latitudes the coriolis force deflects wind to a greater extent than in the tropics, winds become much more zonal (flow parallel to lines of latitude). Also in contrast to the persistent circulation of the tropics, the mid-latitude circulations are quite transient. There are large temperature contrasts, and temperature may vary abruptly over relatively short distances (frontal zones). In these regions of large temperature contrast, potential energy is frequently released and converted into kinetic energy as wind. Near the surface there are many closed pressure systems—cyclones and anticyclones, which are quite mobile, causing frequent changes in weather at any given location. In contrast to the systems near the earth's surface, the motions aloft (above about 3 km) have few closed centers and are mostly waves moving from west to east. The core where speeds are highest in this zonal flow is the jet stream at about 11–14 km aboveground (Fig. 5.27). Where the jet stream undergoes acceleration, divergence occurs at the altitude of the jet stream. This, in turn, promotes convergence near the surface and encourages cyclogenesis (formation of cyclonic motion). Deceleration of the jet stream conversely causes convergence aloft and subsidence near the surface, intensifying high-pressure systems. The strength of the zonal flow is determined by the zonal index, which is the difference in average pressure of two latitude circles such as 35° and 55°. A high index thus represents strong zonal flow; a low index indicates weak zonal flow. A low index is frequently accompanied by closed circulations which provide a greater degree of meridional flow. In keeping with the transient behavior of the mid-latitude circulation, the zonal index varies irregularly, cycling from low to high in periods ranging from 20 to 60 days.

The jet stream is caused by strong temperature gradients, so it is not surprising that it is frequently above the polar front, which lies in the convergence zone between the mid-latitude loop of the general circulation and the

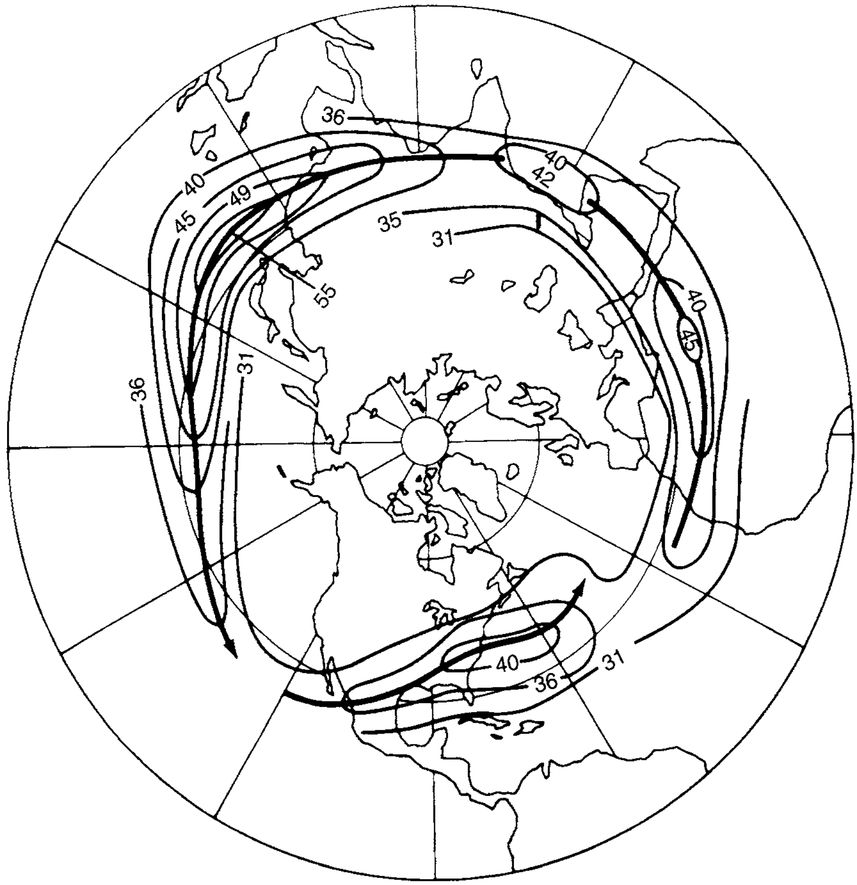

Fig. 5.27. Average position and strength of the jet stream in January between 11 and 14 km above the earth's surface (speeds are in m s^{-1}). *Source*: After Battan [1].

loop nearest the poles (Fig. 5.25). The positions of both the polar front and the jet stream are quite variable, shifting poleward with surface surges of warm air and moving toward the equator with outbreaks of cold air.

C. Polar Region

The circulation cells nearest the poles include rising air along the polar front, movement toward the poles aloft, sinking in the polar regions causing subsidence inversions, and flow toward the equator near the earth's surface. These motions contribute to the heat balance as the moisture in the air rising over the polar front condenses, releasing the heat that was used to evaporate the water nearer the equator. Also, the equatorward-moving air is cold and will be warmed as it is moved toward the tropics.

D. Other Factors

Of considerable usefulness in transporting heat toward the poles are the ocean currents. They are particularly effective because of the high heat content of water. Significant poleward-moving currents are the Brazil, Kuroshio, and Gulf Stream currents. Currents returning cold water toward the equator are the Peru and California currents.

The pressure pattern changes from winter to summer in response to temperature changes. Because most of the Southern Hemisphere consists of ocean, the summer-to-winter temperature differences are moderated. However, the increased landmass in the Northern Hemisphere allows high continental temperatures in summer, causing small equator-to-pole temperature differences; cooling over the continents in winter produces more significant equator-to-pole temperature differences, increasing the westerly winds in general and the jet stream in particular.

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QUESTIONS

1. Verify the intensity of the energy flux from the sun in cal cm^{-2} min^{-1} reaching the outer atmosphere of the earth from the total solar flux of 5.6×10^{27} cal min^{-1} and the fact that the earth is 1.5×10^8 km from the sun. (The surface area of a sphere of radius r is $4\pi r^2$.)
2. Compare the difference in incoming radiation on a horizontal surface at noon on June 22 with that at noon on December 21 at a point at 23.5°N latitude.
3. What is the zenith angle at 1000 local time on May 21 at a latitude of 36°N?
4. At what local time is sunset on August 21 at 40°S latitude?
5. Show the net heating of the atmosphere, on an annual basis, by determining the difference between heat entering the atmosphere and heat radiating to the earth's surface and to space (see Fig. 5.4).
6. If the universal gas constant is 8.31×10^{-2} mb m^{-3} (g mole)$^{-1}$ K^{-1} and the gram molecular weight of dry air is 28.9, what is the mass of a cubic meter of air at a temperature of 293 K and an atmospheric pressure of 996 mb?
7. On a particular day, temperature can be considered to vary linearly with height between 28°C at 100 m aboveground and 26°C at 500 m aboveground. Do you consider the layer between 100 and 500 m aboveground to be stable or unstable?
8. What is the potential temperature of air having a temperature of 288 K at a pressure of 890 mb?
9. Using Wien's displacement law, determine the mean effective temperature of the earth–atmosphere system if the resulting longwave radiation peaks at 11 μm. Contrast the magnitude of the radiant flux at 11 μm with that at 50 μm.
10. What accompanies horizontal divergence near the earth's surface? What effect is this likely to have on the thermal stability of this layer?
11. At what time of day and under what meteorological conditions is maximum ground-level pollution likely to occur at locations several kilometers inland from a shoreline industrial complex whose pollutants are released primarily from stacks of moderate height (about 40–130 m)?
12. When the regional winds are light, at what time of day and what location might high ground-level concentrations of pollutants occur from the low-level sources (less than 20 m) of a town in a north–south-oriented valley whose floor slopes down to the north? Can you answer this question for sources releasing in the range 50–70 m aboveground?
13. Consider the air movement in the Northern Hemisphere (See Fig. 5-27). Explain how persistent organic compounds, such as polychlorinated biphenyls (PCBs) might be found in the tissues of polar bears and in human mother's milk in the arctic regions.
14. New York City is located in a complex geographic and meteorological situation. Explain how local wind systems and heat balances may account for air pollution in and around the city.