Vertical Layers of the Atmosphere

Though people function primarily in the lowest level of the atmosphere, there are times, such as when we fly in aircraft or visit mountainous terrain, when we leave our normal altitude. The thinner atmosphere at these higher altitudes may affect us if we are not accustomed to it. Visitors to Inca ruins in the Andes or high-altitude Himalayan climbers may experience altitude sickness, and even skiers in the Rockies near mile-high Denver may need time to adjust. The air at these levels is much thinner than most of us are used to; by this we mean there is more empty space between air molecules and thus less oxygen and other gases in each breath of air.

As one travels from the surface to outer space, the atmosphere undergoes various changes, and it is necessary to look at the vertical layers that exist with Earth’s atmospheric envelope. There are several systems used to divide the atmosphere into vertical layers. One system uses temperature and rates of temperature changes. Another uses the changes in the content of the gases in the atmosphere, and yet a third deals with the functions of these various layers.

System of Layering by Temperature Characteristics

The atmosphere can be divided into four layers according to differences in temperature and rates of temperature change (Fig. 4.8). The first of these layers, lying closest to Earth’s surface, is the troposphere (from Greek: tropo, turn—the turning or mixing zone), which extends about 8–16 kilometers (5–10 mi) above Earth. Its thickness, which tends to vary seasonally, is least at the poles and greatest at the equator. It is within the troposphere that people live and work, plants grow, and virtually all Earth’s weather and climate take place.

The troposphere has two distinct characteristics that differentiate it from other layers of the atmosphere. One is that the water vapor and particulates of the atmosphere are concentrated in this one layer; they are rarely found in the atmospheric layers above the troposphere. The other characteristic of this layer is that temperature normally decreases with increased altitude. The average rate at which temperatures within the troposphere decrease with altitude is called the normal lapse rate (or the environmental lapse rate); it amounts to 6.5°C per 1000 meters (3.6°F/1000 ft).

The altitude at which the temperature ceases to drop with increased altitude is called the tropopause. It is the boundary that separates the troposphere from the stratosphere—the second layer of the atmosphere. The temperature of the lower part of the stratosphere remains fairly constant (about –57°C, or –70°F) to an altitude of about 32 kilometers (20 mi). It is in the stratosphere that we find the ozone layer that does so much to protect life on Earth from the sun’s UV radiation. As the ozone layer absorbs UV radiation, this absorbed energy results in the release of heat, and thus temperatures increase in the upper parts of the stratosphere. Some water is available in the stratosphere, but it appears as stratospheric ice clouds. These thin veils of ice clouds have no effect on weather as we experience it. Temperatures at the stratopause (another boundary), which is about 50 kilometers (30 mi) above Earth, are about the same as temperatures found on Earth’s surface, although little of that heat can be transferred because the air is so thin.

Above the stratopause is the mesosphere, in which temperatures tend to drop with increased altitude; the mesopause (the last boundary) separates the mesosphere from the thermosphere, where temperatures increase until they approach 1100°C (2000°F) at noon. Again, the air is so thin at this altitude that there is practically a vacuum and little heat can be transferred.

System of Layering by Functional Characteristics

Astronomers, geographers, and communications experts sometimes use a different method of layering the atmosphere, one based on the protective function these layers provide. In this system, the atmosphere is divided into two distinct layers, the lowest of which is the ozonosphere. This layer lies approximately between 15 and 50 kilometers (10–30 mi) above the surface. The ozonosphere is another name for the ozone layer mentioned previously. Again, ozone effectively filters the UV energy from the sun and gives off heat energy instead. As
we have noted, although ozone is a toxic pollutant at Earth’s surface, aloft, it serves a vital function for Earth’s life systems. From about 60–400 kilometers (40–250 mi) above the surface lies the layer known as the ionosphere. This name denotes the ionization of molecules and atoms that occurs in this layer, mostly as a result of UV rays, X-rays, and gamma radiation. Ionization refers to the process whereby atoms are changed to ions through the removal or addition of electrons, giving them an electrical charge. The ionosphere in turn helps shield Earth from the harmful shortwave forms of radiation. This electrically charged layer also aids in transmitting communication and broadcast signals to distant regions of Earth. It is in the ionosphere that the auroras (defined in Chapter 3) occur. The ionosphere gradually gives way to interplanetary space (Fig. 4.9).

**System of Layering by Chemical Composition**

Atmospheric chemists and physicists are at times concerned with the actual chemical makeup of the atmosphere. To this end, there is one more system to divide the atmosphere into two vertical layers. The first is termed the homosphere (from Greek: homo, same throughout). This layer begins on the surface and extends to an altitude of 80 kilometers (50 mi). In this layer, the gases in our atmosphere maintain the same percent by volume as those listed in Table 4.1. There are a few areas of concentration of specific gases, like the water vapor near Earth’s surface and the ozone layer aloft, but for the most part the mixture is homogeneous. Other than rapid decreases in pressure and density while ascending through this layer, this is essentially the same air that we breathe on the surface.

From an altitude of about 80 kilometers (50 mi) and reaching into the vacuum of outer space lies the heterosphere (from Greek: hetero, different). In this layer, atmospheric gases are no longer evenly mixed but begin to separate out into distinct sublayers of concentration. This separation of gases is caused by Earth’s gravity in which heavier gases are pulled closer to the surface and the lighter gases drift farther outward. The regions of concentration and their corresponding gases occur in the following order: nitrogen gas ($N_2$) is the heaviest and therefore the lowermost, followed by atomic oxygen (O), then by helium (He), and finally atomic hydrogen (H)—the lightest element that concentrates at the outermost region (Fig. 4.10).

It is interesting to note that, when watching television news reports from the International Space Station or NASA shuttle missions, the astronauts who we see are still in the atmosphere. The background appears black, making it look like interplanetary space, but in reality these missions still take place within the realm of the outer atmosphere. One should also keep in mind that these different layering systems can focus on the same regions. For example, note that the thermosphere, the ionosphere, and the heterosphere all occupy the same altitudes above Earth—that is, from 80 kilometers (50 mi) and outward. The names are different because of the criteria used in the differing systems.

**Effects of the Atmosphere on Solar Radiation**

As the sun’s energy passes through Earth’s atmosphere, more than half of its intensity is lost through various processes. In addition, as discussed in Chapter 3, the amount of insolation actually received at a particular location depends on not only the processes involved
when water vapor we now examine. Surface is ultimately returned to the atmosphere by processes that Earth’s energy budget is in equilibrium, the 47% received at the retained in the atmosphere, and 34% is returned to space. Because the incoming solar radiation eventually reaches the surface, 19% is affect the other elements. As a result, there is growing concern that one of the elements, human activity, will cause the atmosphere to absorb more Earth-emitted energy, thus raising global temperatures.

Environmental Systems: Earth’s radiation budget

From one year to the next, Earth’s overall average temperature varies very little. This fact indicates that a long-term global balance, or equilibrium, must exist between the energy received and emitted by the Earth system. Note in the diagram that only 47% of the incoming solar energy reaches and is absorbed by Earth’s surface. Eventually, all the energy gained by the atmosphere is lost to space. However, the radiation budget is a dynamic one. In other words, alterations to one element affect the other elements. As a result, there is growing concern that one of the elements, human activity, will cause the atmosphere to absorb more Earth-emitted energy, thus raising global temperatures.

Water as Heat Energy

As it penetrates our atmosphere, some of the incoming solar radiation is involved in several energy exchanges. One such exchange involves how water in the Earth system is altered from one state to another. Water is the only substance that can exist in all three states of matter—as a solid, a liquid, and a gas—within the normal temperature range of Earth’s surface. In the atmosphere, water exists as a clear, odorless gas called water vapor. It is also a liquid in the atmosphere (as clouds, fog, and rain), in the oceans, and in other water bodies on and beneath the surface of Earth. Liquid water is also found within vegetation and animals. Finally, water can be found as a solid in snow and ice in the atmosphere, as well as on and under the surface of the colder parts of Earth.

Not only does water exist in all three states of matter, but it can change from one state to another, as illustrated in Figure 4.12. In doing so, it becomes involved in the heat energy system of Earth. The molecules of a gas move faster than do those of a liquid. Thus, during the process of condensation, when water vapor changes to liquid water, its molecules slow down and some of their
When we look to the skies, we are used to seeing certain colors: the brilliant blue of a clear day, the vibrant reds and oranges of a sunset, clouds ranging from pure white to ominous gray-black, and sometimes even a colorful rainbow. These are wondrous colors, and we sometimes take them for granted. Why they exist, however, can be explained by the concept of atmospheric scattering. This process is explained in the text, but let's examine how it affects the colors of our atmosphere.

It is important to first realize that as sunlight is scattered in our atmosphere, certain wavelengths are scattered more than others. Each wavelength of visible light corresponds to one of the colors of the visible light spectrum. For example, a wavelength of 0.45–0.50 micrometers (1 micrometer = 1/1000 of a millimeter) corresponds to the color blue, and 0.66–0.70 micrometers to the color red. The wavelengths that are scattered, then enter the human eye, are transmitted to the human brain, and then translated into the colors. In other words, the wavelengths that are scattered are the colors that we see.

In 1871, Lord Rayleigh, a radiation scientist, explained that our atmospheric gases tend to scatter the shorter wavelengths of visible light. In other words, Rayleigh scattering makes a clear sky appear blue. In 1908, another radiation specialist, Gustav Mie, explained that oblique (or sharp) sun angles coming through the atmosphere tend to cause the scattering of the longer wavelengths of visible light. In other words, Mie scattering explains why sunsets bring about a red sky. Another form of scattering, called indiscriminant scattering, is caused by tiny water droplets and ice crystals in our atmosphere that tend to scatter all the wavelengths of visible light with equal intensity. When all the wavelengths are scattered equally, white light emerges, and this explains the white clouds we see. (Keep in mind that all clouds are white. When we see dark gray to black storm clouds, we are actually seeing white clouds plus shadows cast by the height and thickness of these massive clouds.)

Rainbows are much more complicated, and their full explanation involves a number of trigonometric functions. Rainbows are caused by water droplets suspended in the atmosphere. The droplets act like tiny glass prisms, which take the incoming white sunlight and project the colors of the spectrum. Key elements necessary to observe a rainbow involve the sun angle, water droplet density and orientation, and the observer angle. When you see a rainbow from a moving vehicle, it is only a matter of time until the angles change and the rainbow disappears.

Incidentally, the colors described above are those distinguishable by the human eye in regard to Earth's atmosphere. Humans visiting another planet with an atmosphere may never see the color blue at all. Furthermore, animals (those that are not color-blind) may see colors other than those described here.

Energy is released into the environment, about 590 calories per gram (cal/g). The molecules of a solid move even more slowly than those of a liquid, so during the process of freezing, when water changes to ice, additional energy is released into the environment, this time 80 calories per gram. When the process is reversed, heat must be added to the ice. Thus, melting ice requires the addition of 80 calories per gram to the ice, from the surrounding environment. Further, evaporation requires the addition of 590 calories per gram to be added to the liquid water, from the environment. This added energy is stored in the water as latent (or hidden) heat. The latent heat of fusion refers to the 80 calories per gram released into the environment when water freezes into ice; to melt ice into water, this heat energy comes from the environment. The latent heat of evaporation (590 cal/g) is added to the water, from the environment, to form water vapor, and the latent heat of condensation (also 590 cal/g) is removed from the water vapor, and released into the environment as it condenses into liquid water. The last is latent heat of sublimation at 670
Heat flows from the sun to Earth. Such systems set into motion by the heating of the Earth back to the atmosphere. This is accomplished through such processes as (1) radiation, (2) conduction, (3) convection (along with the related phenomenon, advection), and (4) the latent heat of condensation (● Fig. 4.13).

**Processes of Heat Energy Transfer**

**Radiation** The process by which electromagnetic energy is transferred from the sun to Earth is called radiation. We should be aware that all objects emit electromagnetic radiation. The characteristics of that radiation depend on the temperature of the radiating body. In general, the warmer the object, the more energy it will emit, and the shorter are the wavelengths of peak emission. Because the sun’s absolute temperature is 20 times that of Earth, we can predict that the sun will emit more energy, and at shorter wavelengths, than Earth. This is borne out by the facts: The energy output per square meter by the sun is approximately 160,000 times that of Earth! Further, most of Earth’s energy is radiated at wavelengths shorter than 4.0 micrometers, whereas most of Earth’s energy is radiated at wavelengths much longer than 4.0 micrometers (see again Fig. 3.9). Thus, **shortwave radiation** from the sun reaches Earth and heats its surface, which, being cooler than the sun, gives off energy in the form of long-waves. It is this **longwave radiation** from Earth’s surface that heats the lower layers of the atmosphere and accounts for the heat of the day.

**Conduction** The means by which heat is transferred from one part of a body to another or between two touching objects is called conduction. Heat flows from the warmer to the cooler part of a body in order to equalize temperature. Conduction actually occurs as heat is passed from one molecule to another in a chainlike fashion. It is conduction that makes the bottom of your soup bowl too hot to touch.

Atmospheric conduction occurs at the interface (zone of contact between) the atmosphere and Earth’s surface. However, it is actually a minor method of heat transfer in terms of warming the atmosphere because it affects only the layers of air closest to the surface. This is because air is a poor conductor of heat. In fact, air is just the opposite of a good conductor; it is a good insulator. This property of air is why a layer of air is sometimes put between two panes of glass to help insulate the window. Air is also used as a layer of insulation in sleeping bags and cold-weather parkas. In fact, if air were a good conductor of heat, our kitchens would become unbearable every time we turned on the stove or oven.

**Convection** In the atmosphere, as pockets of air near the surface are heated, they expand in volume, become less dense than the surrounding air, and therefore rise. This vertical transfer of heat through the atmosphere is called convection; it is the same type of process by which boiling water circulates in a pot on the stove. The water near the bottom is heated first, becoming lighter and less dense as it is heated. As this water tends to rise, colder, denser surface water flows down to replace it. As this new water is warmed, it too flows upward while additional colder water moves downward. This movement within the fluid is called a convective current. These currents set into motion by the heating of a fluid (liquid or gas) make up a convectional system. Such systems account for much of the vertical transfer of heat within the atmosphere and the oceans and are a major cause of clouds and precipitation.
Advection  Advection is the term applied to horizontal heat transfer. There are two major advection agents within the Earth–atmosphere system: winds and ocean currents. Both agents help transfer energy horizontally between the equatorial and polar regions, thus maintaining the energy balance in the Earth–atmosphere system ( Fig. 4.14).

Latent Heat of Condensation  As we have seen, when water evaporates, a significant amount of energy is stored in the water vapor as latent heat (see again Fig. 4.12). This water vapor is then transported by advection or convection to new locations where condensation takes place and the stored energy is released. This process plays a major role in the transfer of energy within the Earth system. The latent heat of evaporation helps cool the atmosphere while the latent heat of condensation helps warm the atmosphere and, in addition, is a source of energy for storm activity. The power of all severe weather is supplied by the latent heat of condensation.

The Heat Energy Budget

The Heat Energy Budget at Earth’s Surface  Now that we know the various means of heat transfer, we are in a position to examine what happens to the 47% of solar energy that reaches Earth’s surface (see again Fig. 4.11). Approximately 14% of this energy is emitted by Earth in the form of longwave radiation. This 14% includes a net loss of 6% (of the total) directly to outer space, and the other 8% is captured by the atmosphere. In addition, there is a net transfer back to the lower atmosphere (by conduction and convection) of 10 of the 47% that reached Earth. The remaining 23% returns to the atmosphere through the release of latent heat of condensation. Thus, 47% of the sun’s original insolation that reached Earth’s surface is all returned to other segments of the system. There has been no long-term gain or loss. Therefore, at Earth’s surface, the heat energy budget is in balance.

Examination of the heat energy budget of Earth’s surface helps us understand the open energy system that is involved in the heating of the atmosphere. The input in the system is the incoming shortwave solar radiation that reaches Earth’s surface; this is balanced by the output of longwave terrestrial (Earth’s) radiation back to the atmosphere and to space. As these functions adjust to remain in balance, we can say that the overall heat budget of Earth’s surface is in a state of dynamic equilibrium.

Of course, it should be noted that the percentages mentioned earlier represent an oversimplification in that they refer to net losses that occur over a long period of time. In the shorter term, heat may be passed from Earth to the atmosphere and then back to Earth in a chain of cycles before it is finally released into space. The absorption and reflection of incoming solar radiation, and the emission of outgoing terrestrial radiation, can all be affected by surface ground cover and any human activity that may change the surface cover.

The Heat Energy Budget in the Atmosphere  At one time or another, about 60% of the solar energy intercepted by the Earth system is temporarily retained by the atmosphere. This includes 19% of direct solar radiation absorbed by the clouds and the ozone layer, 8% emitted by longwave radiation from the Earth’s surface, 10% transferred from the surface by conduction and convection, and 23% released by the latent heat of condensation. Some of this energy is recycled back to the surface for short periods of time, but eventually all of it is lost into outer space as more solar energy is received. Hence, just as was the case at Earth’s surface, the heat energy budget in the atmosphere is in balance over long periods of time.
periods of time—a dynamically stable system. However, many scientists believe that an imbalance in the heat energy budget, with possible negative effects, could develop due to the greenhouse effect.

Variations in the Heat Energy Budget  Remember that the figures we have seen for the heat energy budget are averages for the whole Earth over many years. For any particular location, the heat energy budget is most likely not balanced. Some places have a surplus of incoming solar energy over outgoing energy loss, and others have a deficit. The main causes of these variations are differences in latitude and seasonal fluctuations.

As we have noted previously, the amount of insolation received is directly related to latitude (see again Fig. 4.14). In the tropical zones, where insolation is high throughout the year, more solar energy is received at Earth’s surface and in the atmosphere than can be emitted back into space. In the Arctic and Antarctic zones, on the other hand, there is so little insolation during the winter, when Earth is still emitting longwave radiation, that there is a large deficit for the year. Locations in the middle-latitude zones have lower deficits or surpluses, but only at about latitude 38° is the budget balanced. If it were not for the heat transfers within the atmosphere and the oceans, the tropical zones would get hotter and the polar zones would get colder through time.

At any location, the heat energy budget varies throughout the year according to the seasons, with a tendency toward a surplus in the summer or high-sun season and a tendency toward a deficit 6 months later. Seasonal differences may be small near the equator, but they are great in the middle-latitude and polar zones.

Air Temperature

Temperature and Heat

Although heat and temperature are highly related, they are not the same. Heat is a form of energy—the total kinetic energy of all the atoms that make up a substance. All substances are made up of molecules that are constantly in motion (vibrating and colliding) and therefore possess kinetic energy—the energy of motion. This energy is manifested as heat. Temperature, on the other hand, is the average kinetic energy of the individual molecules of a substance. When something is heated, its atoms vibrate faster, and its temperature increases. It is important to remember that the amount of heat energy depends on the mass of the substance under discussion, whereas the temperature refers to the energy of individual molecules. Thus, a burning match has a high temperature but minimal heat energy; the oceans have moderate temperatures but high heat energy content.

Temperature Scales

Three different scales are generally used in measuring temperature. The one with which Americans are most familiar is the Fahrenheit scale, devised in 1714 by Daniel Fahrenheit, a German scientist. On this scale, the temperature at which water boils at sea level is 212°F, and the temperature at which water freezes is 32°F. This scale is used in the English system of measurements.

The Celsius scale (also called the centigrade scale) was devised in 1742 by Anders Celsius, a Swedish astronomer. It is part of the metric system. The temperature at which water freezes at sea level on this scale was arbitrarily set at 0°C, and the temperature at which water boils was designated as 100°C.

The Celsius scale is used nearly everywhere but in the United States, and even in the United States, the Celsius scale is used by the majority of the scientific community. By this time, you have undoubtedly noticed that throughout this book comparable figures in both the Celsius and Fahrenheit scales are given side by side for temperatures. Similarly, whenever important figures for distance, area, weight, or speed are given, we use the metric system followed by the English system. Appendix A at the back of your text may be used for comparison and conversion between the two systems.

Figure 4.15 can help you compare the Fahrenheit and Celsius systems as you encounter temperature figures outside this

![Figure 4.15](image-url)
book. In addition, the following formulas can be used for conversion from Fahrenheit to Celsius or vice versa:

\[ C = (F - 32) ÷ 1.8 \]
\[ F = (C ÷ 1.8) + 32 \]

The third temperature scale, used primarily by scientists, is the Kelvin scale. Lord Kelvin, a British radiation scientist, felt that negative temperatures were not proper and should not be used. In his mind, no temperature should ever go below zero. His scale is based on the fact that the temperature of a gas is related to the molecular movement within the gas. As the temperature of a gas is reduced, the molecular motion within the gas slows. There is a temperature at which all molecular motion stops and no further cooling is possible. This temperature, approximately \(-273°C\), is termed absolute zero. The Kelvin scale uses absolute zero as its starting point. Thus, \(0^K\) equals \(-273°C\). Conversion of Celsius to Kelvin is expressed by the following formula:

\[ K = C + 273 \]

**Short-Term Variations in Temperature**

Local changes in atmospheric temperature can have a number of causes. These are related to the mechanics of the receipt and dissipation of energy from the sun and to various properties of Earth's surface and the atmosphere.

**The Daily Effects of Insolation** As we noted earlier, the amount of insolation at any particular location varies both throughout the year (annually) and throughout the day (diurnally). Annual fluctuations are associated with the sun’s changing declination and hence with the seasons. Diurnal changes are related to the rotation of Earth about its axis. Each day, insolation receipt begins at sunrise, reaches its maximum at noon (local solar time), and returns to zero at sunset.

Although insolation is greatest at noon, you may have noticed that temperatures usually do not reach their maximum until 2–4 p.m. (Fig. 4.16). This is because the insolation received by Earth from sunrise until the afternoon hours exceeds the energy being lost through Earth radiation. Hence, during that period, as Earth and atmosphere continue to gain energy, temperatures normally show a gradual increase. Sometime around 3–4 p.m., when outgoing Earth radiation begins to exceed insolation, temperatures start to fall. The daily lag of Earth radiation and temperature behind insolation is accounted for by the time it takes for Earth’s surface to be heated to its maximum and for this energy to be radiated to the atmosphere.

Insolation receipt ends at sunset, but energy that has been stored in Earth’s surface layer during the day continues to be lost throughout the night and the ability to heat the atmosphere decreases. The lowest temperatures occur around dawn, when the maximum amount of energy has been emitted and before replenishment from the sun can occur. Thus, if we disregard other factors for the moment, we can see that there is a predictable hourly change in temperature called the **daily march of temperature**. There is a gentle decline from midafternoon until dawn and a rapid increase in the 8 hours or so from dawn until the next maximum is reached.

**Cloud Cover** The extent of cloud cover is another factor that affects the temperature of Earth’s surface and the atmosphere (Fig. 4.17). Weather satellites have shown that, at any time, about 50% of Earth is covered by clouds (Fig. 4.18). This is important because a heavy cloud cover can reduce the amount of insolation a place receives, thereby causing daytime temperatures to be lower on a cloudy day. On the other hand, we also have the greenhouse effect, in which clouds, composed in large part of water droplets, are capable of absorbing heat energy radiating from Earth. Clouds therefore keep temperatures near Earth’s surface warmer than they would otherwise be, especially at night. The general effect of cloud cover, then, is to moderate temperature by lowering the potential maximum and raising the potential minimum temperatures. In other words, cloud cover makes for cooler days and warmer nights.

**Differential Heating of Land and Water** For reasons we will later explain in detail, bodies of water heat and cool more slowly than the land. The air above Earth’s surface is heated or cooled in part by what is beneath it. Therefore, temperatures over bodies of water or on land subjected to ocean winds (maritime locations) tend to be more moderate than those of land-locked places at the same latitude. Thus, the greater the **continentality** of a location (the distance removed from a large body of water), the less its temperature pattern will be modified.
**FIGURE 4.17**
The effect of cloud cover on temperatures. (a) By intercepting insolation, clouds produce lower air temperatures during the day. (b) By trapping longwave radiation from Earth, clouds increase air temperatures at night. The overall effect is a great reduction in the diurnal temperature range.

Why do desert regions have large diurnal variations in temperature?

**FIGURE 4.18**
This composite of several satellite images shows a variety of cloud cover and storm systems across Earth.

In general which are the cloudiest latitude zones and which are the zones with the clearest skies?
Reflection. The capacity of a surface to reflect the sun’s energy is called its albedo; a surface with a high albedo has a high percentage of reflection. The more solar energy that is reflected back into space by Earth’s surface, the less that is absorbed for heating the atmosphere. Temperatures will be higher at a given location if its surface has a low albedo rather than a high albedo.

As you may know from experience, snow and ice are good reflectors; they have an albedo of 90–95%. Or put another way, only 5–10% of the incoming solar radiation is absorbed by the snow and ice. This is one reason why glaciers on high mountains do not melt away in the summer or why there may still be snow on the ground on a sunny day in the spring. Most of the solar energy is reflected away. A forest, on the other hand, has an albedo of only 10–15% (or 85–90% absorption), which is good for the trees because they need solar energy for photosynthesis. The albedo of cloud cover varies, from 40 to 80%, according to the thickness of the clouds. The high albedo of many clouds is why much solar radiation is reflected directly back into space by the atmosphere.

The albedo of water varies greatly, depending on the depth of the water body and the angle of the sun’s rays. If the angle of the sun’s rays is high, smooth water will reflect little. In fact, if the sun is vertical over a calm ocean, the albedo will be only about 2%. However, a low sun angle, such as just before sunset, causes an albedo of more than 90% from the same ocean surface (Fig. 4.19). Likewise, a snow surface in winter, when solar angles are lower, can reflect up to 95% of the energy striking it, and skiers must constantly be aware of the danger of severe sunburns and possible snow blindness from reflected solar radiation. In a similar fashion, the high albedo of sand causes the sides of sunbathers’ legs to burn faster when they lie on the beach.

Horizontal Air Movement. We have already seen that advection is the major mode of horizontal transfer of heat and energy over Earth’s surface. Any movement of air due to the wind, whether on a large or small scale, can have a significant short-term effect on the temperatures of a given location. Thus, wind blowing from an ocean to land will generally bring cooler temperatures in summer and warmer temperatures in winter. Large quantities of air moving from polar regions into the middle latitudes can cause sharp drops in temperature, whereas air moving poleward will usually bring warmer temperatures.

Vertical Distribution of Temperature. Normal Lapse Rates. We have learned that Earth’s atmosphere is primarily heated from the ground up as a result of longwave terrestrial radiation, conduction, and convection. Thus, temperatures in the troposphere are usually highest at ground level and decrease with increasing altitude. As noted earlier in the chapter, this decrease in the free air of approximately 6.5°C per 1000 meters (3.6°F/1000 ft) is known as the normal lapse rate.

The lapse rate at a particular place can vary for a variety of reasons. Lower lapse rates can exist if denser and colder air is drained into a valley from a higher elevation or if advectional winds bring air in from a cooler region at the same altitude. In each case, the surface is cooled. On the other hand, if the surface is heated strongly by the sun’s rays on a hot summer afternoon, the air near Earth will be disproportionately warm, and the lapse rate will increase. Fluctuations in lapse rates due to abnormal temperature conditions at various altitudes can play an important role in the weather a place may have on a given day.

Temperature Inversions. Under certain circumstances, the normal observed decrease of temperature with increased altitude might be reversed; temperature may actually increase for several hundred meters. This is called a temperature inversion.

Some inversions take place 1000 or 2000 meters above the surface of Earth where a layer of warmer air interrupts the normal decrease in temperature with altitude (Fig. 4.20). Such inversions tend to stabilize the air, causing less turbulence and discouraging both precipitation and the development of storms. Upper air inversions may occur when air settles slowly from the upper atmosphere. Such air is compressed as it sinks and rises in temperature, becoming more stable and less buoyant. Inversions caused by descending air are common at about 30–35° north and south latitudes.

An upper air inversion common to the coastal area of California results when cool marine air blowing in from the Pacific Ocean moves under stable, warmer, and lighter air aloft created by subsidence and compression. Such an inversion layer tends to maintain itself; that is, the cold underlying air is heavier and cannot rise above the warmer air above. Not only does the cold air resist rising or moving, but pollutants, such as smoke, dust particles, and automobile exhaust, created at Earth’s surface, also fail to disperse. They therefore accumulate in the lower atmosphere. This situation is particularly acute in the Los Angeles area, which is a basin surrounded by higher mountainous areas (Fig. 4.21). Cooler air blows into the basin from the ocean and then cannot escape either horizontally, because of the landform barriers, or vertically, because of the inversion.

Some of the most noticeable temperature inversions are those that occur near the surface when Earth cools the lowest layer of air through conduction and radiation (Fig. 4.22). In this situation, the
The coldest air is nearest the surface and the temperature rises with altitude. Inversions near the surface most often occur on clear nights in the middle latitudes. They may be enhanced by snow cover or the recent advection of cool, dry air into the area. Such conditions produce extremely rapid cooling of Earth’s surface at night as it loses the day’s insolation through radiation. Then the layers of the atmosphere that are closest to Earth are cooled by radiation and conduction more than those at higher altitudes. Calm air conditions near the surface help produce and partially result from these temperature inversions.

**Surface Inversions: Fog and Frost** Fog and frost will be discussed again in Chapter 6, but they often occur as the result of a surface inversion. Especially where Earth’s surface is hilly, cold, dense surface air will tend to flow downslope and accumulate in the lower valleys. The colder air on the valley floors and other low-lying areas sometimes produces fog or, in more extreme cases, a killing frost. Farmers use a variety of methods to prevent such frosts from destroying their crops. For example, fruit trees in California are often

![Figure 4.20](image1.png)

*FIGURE 4.20* (Left) Temperature inversion caused by subsidence of air. (Right) Lapse rate associated with the column of air (A) in the left-hand drawing. Why is the pattern (to the right) called a temperature inversion?

![Figure 4.21](image2.png)

*FIGURE 4.21* Conditions producing smog-trapping inversion in the Los Angeles area. Why is the air clear above the inversion?

![Figure 4.22](image3.png)

*FIGURE 4.22* Temperature inversion caused by the rapid cooling of the air above the cold surface of Earth at night. What is the significance of an inversion?
planted on the warmer hillsides instead of in the valleys. Farmers may also put blankets of straw, cloth, or some other insulator over their plants. This prevents the escape of Earth’s heat radiation to outer space and thereby keeps the plants warmer. Large fans and helicopters are sometimes used in an effort to mix surface layers and disturb the inversion (Fig. 4.23a). Huge orchard heaters that warm the air can also be used to disturb the temperature layers. Smudge pots, an older method of preventing frost, pour smoke into the air, which provides a blanket of insulation much like blankets of cloth or straw (Fig. 4.23b). However, smudge pots have declined in favor because of their air-pollution potential.

**Controls of Earth’s Surface Temperatures**

Variations in temperatures over Earth’s surface are caused by several controls. The major controls are (1) latitude, (2) land and water distribution, (3) ocean currents, (4) altitude, (5) landform barriers, and (6) human activity.

**Latitude** Latitude is the most important control of temperature variation involved in weather and climate. Recall that, because of the inclination and parallelism of Earth’s axis as it revolves around the sun (Chapter 3), there are distinct patterns in the latitudinal distribution of the seasonal and annual receipt of solar energy over Earth’s surface. This has a direct effect on temperatures. In general, annual insolation tends to decrease from lower latitudes to higher latitudes (see again Fig. 3.16). Table 4.2 shows the average annual temperatures for several locations in the Northern Hemisphere. We can see that, responding to insolation (with one exception), a poleward decrease in temperature is true for these locations. The exception is near the equator itself. Because of the heavy cloud cover in equatorial regions, annual temperatures there tend to be lower than at places slightly to the north or south, where skies are clearer.

Another very simple way to see this general trend of decreasing temperatures as we move toward the poles is to think about the kinds of clothes we would take along for 1 month—say, January—if we were to visit Ciudad Bolivar, Venezuela; Raleigh, North Carolina; or Point Barrow, Alaska (see Table 4.2).

**Land and Water Distribution** Not only do the oceans and seas of Earth serve as storehouses of water for the whole system, but they also store tremendous amounts of heat energy. Their widespread distribution makes them an important atmospheric control that does much to modify the atmospheric elements. All things heat and cool at different rates. This is especially true...