3

Atmospheric temperatures

In Chapter 2 we considered some of the chemical changes that occur when solar radiation is absorbed by the atmosphere of a planet. Chemical changes are only a part of the story, however. The energy carried by solar radiation affects the temperature of an atmosphere as well as the chemical composition.

The heating effect of solar radiation can be seen quite readily in Fig. 3–1, which shows how temperature varies with altitude at middle latitudes in the Earth's atmosphere. There are three maxima in the temperature profile; the first is at the ground, where the temperature is about 290°K; the second is at the level called the *stratopause*, at a height of 50 km, where the temperature is about 280°K; and the third occurs at heights above 200 km, in the layer of the atmosphere called the *thermosphere*, where the temperature can rise to 1000°K or more. On the right of the figure there is an indication of the height in the atmosphere at which solar radiation of various wavelengths undergoes absorption. This is a simplified representation of Fig. 2–4, which shows the altitude of absorption plotted against wavelength.

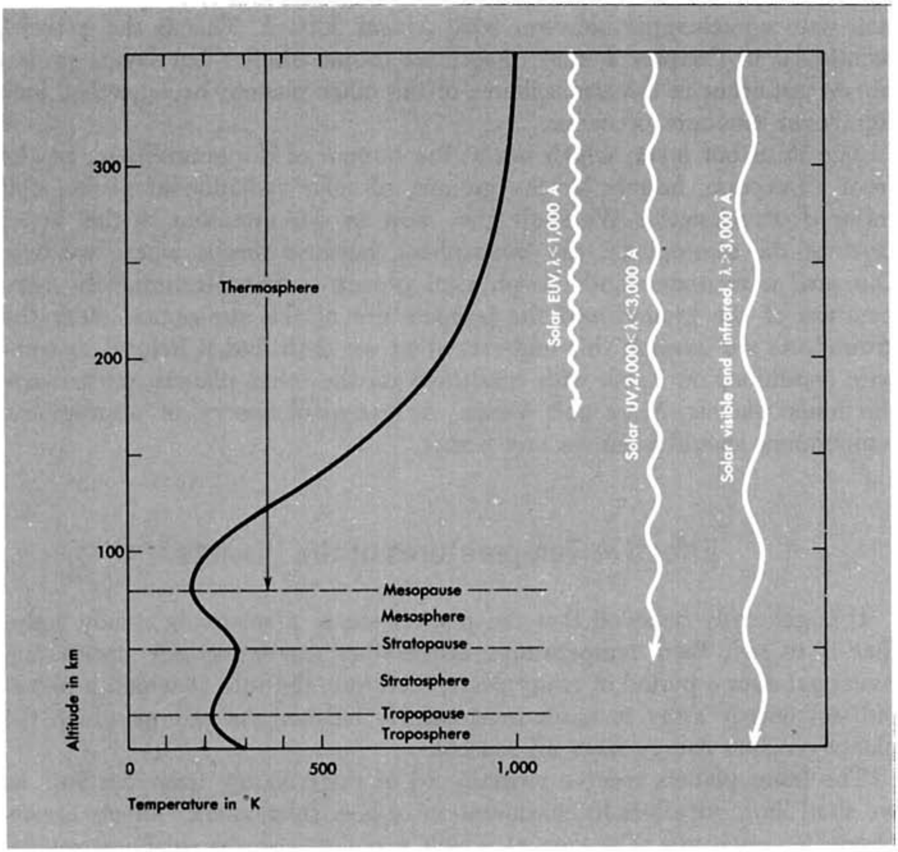

FIGURE 3-1 *The variation of temperature with altitude at middle latitudes in the Earth's atmosphere. Layers of the atmosphere are called troposphere, stratosphere, mesosphere, and thermosphere; these are separated by features called tropopause, stratopause, and mesopause. The high temperature regions of the atmosphere result from the absorption of solar radiation at different wavelengths, as shown on the right.*

The high temperature in the thermosphere is caused by solar heat, deposited at great heights as a result of the absorption of extreme ultraviolet radiation. The absorption process is photoionization, which we discussed in Chapter 2 as the source of the ionosphere. Similar processes of photoionization occur high in the atmospheres of all the planets, and we may anticipate that relatively hot thermospheres are a feature of all planetary atmospheres.

The second hot layer in the Earth's atmosphere, at a height of about 50 km, is the result of the absorption, by ozone, of solar ultraviolet radiation with wavelengths between 2000 Å and 3000 Å. This is the process, mentioned in Chapter 2, that dissociates ozone. Similar hot layers probably do not occur in the atmospheres of the other planets, because they lack significant amounts of ozone.

The third hot layer, which lies at the bottom of the atmosphere, results from absorption, largely by the ground, of solar radiation at visible and infrared wavelengths. We shall give most of our attention to this lower layer of the atmosphere, the *troposphere*, because this is where we live. Our goal is to understand the physical processes that determine the temperature of the ground and the temperature of the atmosphere near the ground. As we develop this understanding we shall find it helpful to compare conditions on Earth with conditions on the other planets, particularly the inner planets, Mars and Venus. A successful theory of atmospheric temperature should work on any planet.

Effective Temperatures of the Planets

It is generally believed that the planets are in a relatively steady state; that is to say, their temperatures are neither increasing nor decreasing. Averaged over a period of many years, therefore, the rate at which a planet radiates energy away to space must exactly balance the rate at which the planet receives energy from all sources.

The inner planets receive virtually all of their energy from the Sun, so we shall limit ourselves to consideration of how solar energy affects atmospheric temperatures. The rate at which a planet absorbs solar energy depends strongly on the distance of the planet from the Sun, because the flux of solar radiation—the amount of energy flowing across unit area in unit time (erg cm^{-2} sec^{-1})—varies inversely as the square of this distance (see Chapter 1). Table 3–1 shows the distance of each planet from the Sun and gives corresponding values of the solar flux. Each planet intercepts solar radiation at a rate (erg sec^{-1}) given by the product of the solar flux (erg cm^{-2} sec^{-1}) and the area of the disk of the planet (cm^2) as seen from the direction of the Sun (Fig. 3–2). This area is π times the square of the radius of the planet.

Not all of the radiation intercepted by a planet is absorbed. A fraction of the incident energy is reflected back to space by clouds and by the surface. This fraction is called the *albedo* of the planet. The albedo is different for every planet, depending on the nature of the atmosphere and surface. Values, determined from astronomical observations, are given in Table 3–1. Since the albedo is the fraction of the incident energy reflected by the planet, the fraction that is absorbed is $(1 - \text{albedo})$.

Table 3–1

Effective Temperatures

Planet	Distance from Sun (10^6 km)	Flux of Solar Radiation (10^6 erg cm^{-2} sec^{-1})	Albedo	T_e (°K)
Mercury	58	9.2	.058	442
Venus	108	2.6	.71	244
Earth	150	1.4	.33	253
Mars	228	.60	.17	216
Jupiter	778	.049	.73	87
Saturn	1430	.015	.76	63
Uranus	2870	.0037	.93	33
Neptune	4500	.0015	.84	32
Pluto	5900	.00089	.14	43

Combining these results, we find that a planet absorbs solar energy at a rate given by the following expression:

$$\text{Energy absorbed} = \pi \times (\text{Radius})^2 \times \text{Solar flux} \times (1 - \text{albedo}) \quad (3\text{–}1)$$

We must now consider the rate at which the planet radiates energy away to space because, as we have already noted, the rate of loss of energy must be equal to the rate of absorption of energy in the long run.

We described in Chapter 1 how hot material radiates energy more rap-

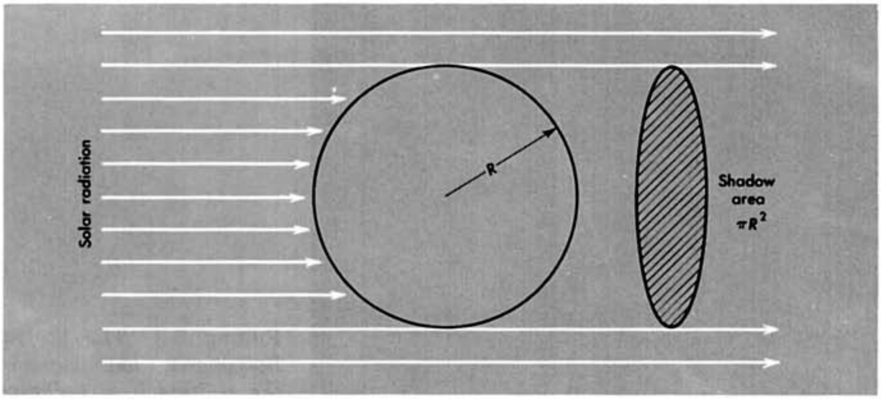

FIGURE 3-2 *From the direction of the Sun a planet looks like a disk with an area π times the square of the radius.*

idly than cold material. According to the *Stefan-Boltzmann law*, the amount of heat energy radiated by a surface of unit area in unit time is proportional to the fourth power of the temperature (Fig. 3–3). We can write this law as

$$\text{Flux (erg cm}^{-2}\text{ sec}^{-1}) = \sigma \times (\text{Temperature})^4 \qquad (3\text{-}2)$$

where σ is the constant of proportionality, called the *Stefan-Boltzmann constant* ($\sigma = 5.67 \times 10^{-5}$ erg cm^{-2} deg^{-4} sec^{-1}). Each unit area of planetary surface radiates energy at a rate given by Eq. (3–2), so to get the total rate at which the planet loses energy to space we must multiply the flux by the total surface area. The surface area of a sphere is 4π times the square of the radius, so we find

$$\text{Energy radiated} = 4\pi \times (\text{Radius})^2 \times \sigma \times (\text{Temperature})^4 \qquad (3\text{–}3)$$

If we now equate the rate at which energy is radiated by the planet (Eq. 3–3) to the rate at which energy is absorbed by the planet (Eq. 3–1) and rearrange terms, we obtain an expression for the temperature

$$\text{Temperature} = \sqrt[4]{\frac{\text{Solar flux} \times (1 - \text{albedo})}{4 \times \sigma}} \qquad (3\text{–}4)$$

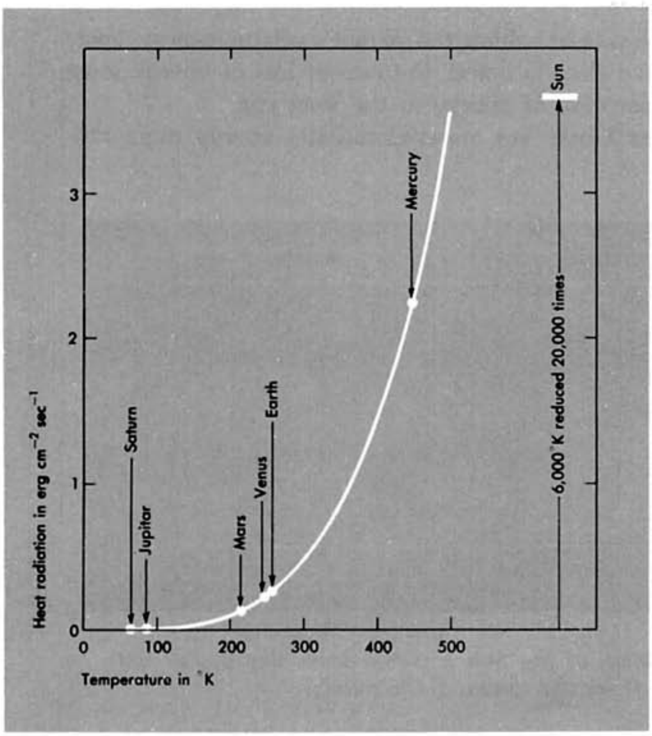

FIGURE 3-3 *The Stefan-Boltzmann law equates the emitted heat radiation to 5.67 × 10*$^{-5}$ *× (Temperature)*^{4}*. Effective temperatures of the Sun and planets are shown.*

The temperature calculated in this manner is called the *effective temperature* (T_e) of a planet. Note that the radius of the planet has cancelled out of the expression, so the effective temperature depends not on the size of the planet but only on the albedo and the distance of the planet from the Sun. Table 3–1 shows calculated values of the effective temperatures of the planets. Mercury, the planet closest to the Sun, has a very high effective temperature. The outer planets are cold because of their enormous distances from the Sun. Venus and Earth have almost the same effective temperature, even though Earth is farther from the Sun because Venus, with its unbroken cloud cover, has a higher albedo than Earth and absorbs a smaller fraction of the solar radiation incident upon it.

Surface Temperatures

An important distinction must be made between the effective temperatures we have calculated and the temperatures of planetary surfaces. If a planet has a substantial atmosphere, the atmosphere can absorb all heat radiation from the lower surface before the radiation penetrates into outer space. Thus, an instrument in space looking at the planet does not detect radiation from the surface. The radiation it "sees" comes from some level higher in the atmosphere.

The effective temperature is the temperature of this emitting region, and lower levels may have much higher temperatures. On Earth, for example, the average temperature of the surface is 288°K, but the effective temperature is only 253°K. The difference is even more striking on Venus. Ground-based measurements of thermal radio waves emitted by the surface of Venus show that the temperature there is about 700°K, close to the melting point of lead. This surprising result has been confirmed by measurements of atmospheric properties made by the Venera and Mariner spacecraft. But the effective temperature of Venus is only 244°K, close to the effective temperature of Earth. For Mars we expect the average surface temperature to be only a little above the effective temperature, 216°K, because the atmosphere of Mars is thin.

The Greenhouse Effect

The extent to which the temperature of the surface and the lower portions of the atmosphere can differ from the effective temperature depends not only on the mass of the atmosphere but also on its constituents. The important property is the opacity of the gas to electromagnetic radiation and particularly to the infrared radiation emitted by the planet. In Fig. 3–4 we show, at each wavelength, the fraction of the light absorbed from a beam passing directly through the Earth's atmosphere. We also show the shape of the spectrum of incident solar radiation, peaking in the visible because of the high effective temperature of the Sun, and the shape of the thermal emission spectrum of the Earth, peaking far out in the infrared. This is in accordance with our discussion of the Planck spectra in Fig. 1–11. According to *Wien's radiation law*, the wavelength of the peak is inversely proportional to the temperature. Because the effective temperature of the Earth is approximately $\frac{1}{24}$ times the effective temperature of the Sun, the peak wavelength of terrestrial radiation is about 24 times the peak wavelength of solar radiation.

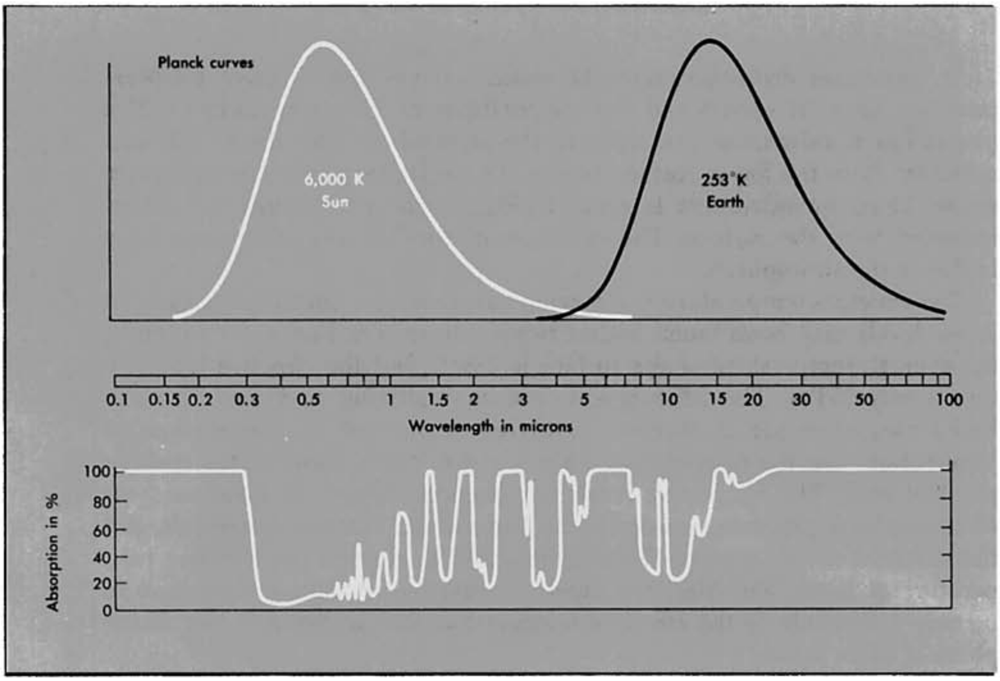

FIGURE 3-4 *At the top are the Planck curves that show the proportion of energy radiated at each wavelength by the Earth and the solar radiation incident on the Earth. There is very little overlap between the curves, which enables us to talk of different atmospheric characteristics for short-wave solar radiation and long-wave heat radiation. At the bottom is the percentage of the radiation at each wavelength that is absorbed in the atmosphere. Absorption is strong on the average for the planetary radiation but weak on the average for the solar radiation. (After R. M. Goody, 1954, Physics of the Stratosphere, Cambridge University Press.)*

We conclude from Fig. 3-4 that the atmosphere is moderately transparent in the visible and that much of the solar radiation can pass right through the atmosphere without being absorbed. On the other hand, minor atmospheric constituents, of which water vapor is the most important, absorb strongly in the infrared so the atmosphere is largely opaque to the planetary heat radiation.

What happens when the atmosphere absorbs radiation emitted from the surface of the planet? The atmosphere cannot steadily accumulate energy or it would become hotter and hotter. Instead, it emits radiation at the same rate as it absorbs. The radiation is reemitted in all directions, and a substantial part of it is intercepted and absorbed by the surface. So the surface of the planet is heated not only by direct sunlight but also by infrared radiation emitted by the atmosphere. For this reason the surface of a planet must radiate away more energy than it receives directly from the Sun, and the surface can have a temperature that exceeds the effective temperature of the planet.

These ideas are given quantitative expression in Fig. 3-5, which shows the mean annual heat budget of the Earth and the atmosphere, derived from meteorological measurements. The left-hand side of the figure shows

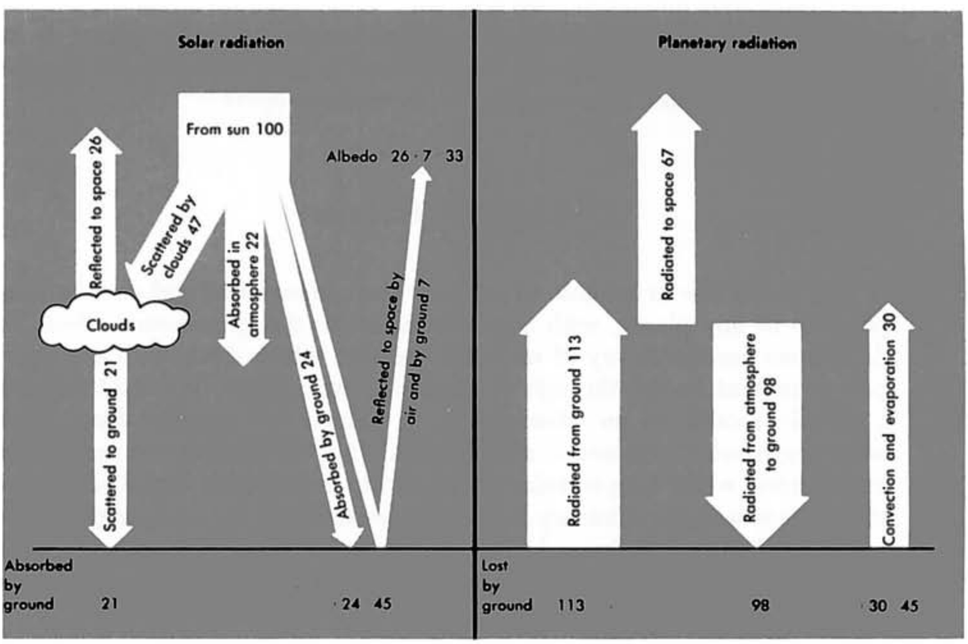

FIGURE 3-5 *The heat budget of the Earth and the atmosphere. (After J. London and K. Sasamori, 1971.)*

what happens to short-wave (visible) solar radiation. All quantities are expressed in units such that the incident solar flux has a magnitude of 100 units. Of this incident flux, about 22 units are absorbed somewhere in the atmosphere and about 33 units are reflected back to space by the ground, the atmosphere, and the clouds. These 33 units represent the Earth's albedo. This energy is lost to the Earth entirely and plays no role in heating either the ground or the atmosphere. The remaining 45 units of the incident radiation are absorbed by the ground. Both the atmosphere and the ground must lose the energy they have absorbed, and they do so by radiating in the long-wave (infrared) region of the spectrum. What happens to this energy is shown on the right of the figure.

The flux of infrared radiation emitted by the ground amounts to 113 of our arbitrary units, two and a half times as much as the flux of solar radiation absorbed by the ground. This extra energy comes, as the figure shows, from the atmosphere. The ground absorbs a long-wave flux from the atmosphere that is equal to 98 units. As a result, the net loss of radiation from the surface in the long-wave region of the spectrum is only 15 units. Since 45 units of solar radiation are absorbed by the ground, we appear to have too much energy going into the ground. The additional ways of removing heat from the ground are shown on the right of Fig. 3–5. Evaporation of water is one way, and heat convection is another. We shall discuss evaporation and convection later on.

This phenomenon, in which the surface temperature of a planet is increased because the atmosphere is translucent to solar radiation but opaque to infrared radiation, is known as the *greenhouse effect.**

Radiative Transfer

We would like to be able to estimate the atmospheric and surface temperature of any planet, with due allowance for the greenhouse effect. To do this we use the theory of *radiative transfer*, which deals with the transport of radiant energy through an absorbing atmosphere. We shall examine a simple model of an atmosphere, in which we assume that short-wavelength solar radiation is absorbed only by the ground and not by the atmosphere, while long-wavelength planetary radiation is absorbed by the atmosphere with an efficiency that is independent of the wavelength of the radiation. Figure 3–4 shows how complete is the separation in wavelength of the solar radiation and the planetary radiation; it is reasonable to assume different optical properties for the two.

*Some writers prefer to avoid this term because the analogy to the domestic greenhouse is not complete, but the term is evocative and rather widely used.

We shall further assume that radiation is the only way of carrying heat from one level of the atmosphere to another. This is called the assumption of *radiative equilibrium*. It has the merit of being the simplest model of heat transfer that has any relevance to our problem, although at a later point we will have to include the effects of convection, condensation, and winds in order to explain observed phenomena. For the present, however, we shall pursue this model to discover its implications.

In order to calculate a radiative equilibrium temperature profile, let us divide the atmosphere into horizontal layers (see Fig. 3–6). The thickness of the layers is adjusted so that radiation emitted in one layer is absorbed in an adjacent layer. The layers must be neither too thick nor too thin. Layers are too thick if radiation is emitted and reabsorbed in the same layer. Layers are too thin if radiation transverses one or more layers before undergoing absorption. Each layer, therefore, is just thick enough to absorb the radiation falling onto it. The mechanism of radiative transfer is one of passing energy from one layer to the next; the radiation emitted from each layer is absorbed by its two nearest neighbors, which in turn emit to their nearest neighbors, and so on (see Fig. 3–6).

The thickness of the layers can generally be expected to increase as the altitude increases because at higher altitudes the density of a thoroughly mixed absorbing atmospheric gas is less, allowing radiation to travel farther before being absorbed. The total number of layers into which an atmosphere can be divided in this manner is called the *optical thickness* of the atmosphere. The optical thickness depends on how much atmosphere there is and also on the efficiency with which atmospheric gases absorb infrared heat radiation.

Let us consider the balance of energy in the topmost layer of the atmosphere (see Fig. 3–6). If the average temperature of the layer is T_1, the layer radiates energy away to space at a rate σT_1^4, and it radiates energy at the same rate downward to the layer below.

First consider the radiation to space. According to our discussion of the effective temperature, on the average the outgoing radiation (σT_1^4) must be equal to the incoming radiation, which in turn is equal to σT_e^4 [see Eq. (3–2)]. From this we conclude that T_e and T_1 are equal.

Now consider the radiation balance of the top layer itself. The only radiation that it can absorb comes from the layer immediately below, at a rate σT_2^4, where T_2 is the average temperature of the second layer from the top. If we now equate the energy radiated by the topmost layer to the energy absorbed, we find

$$\sigma T_1^4 \text{ (upward)} + \sigma T_1^4 \text{ (downward)} = \sigma T_2^4 \text{ (upward)} \qquad (3\text{–}5)$$

This gives us a relationship between the temperatures in the top two layers

$$T_2^4 = 2\, T_1^4 = 2\, T_e^4 \qquad (3\text{–}6)$$

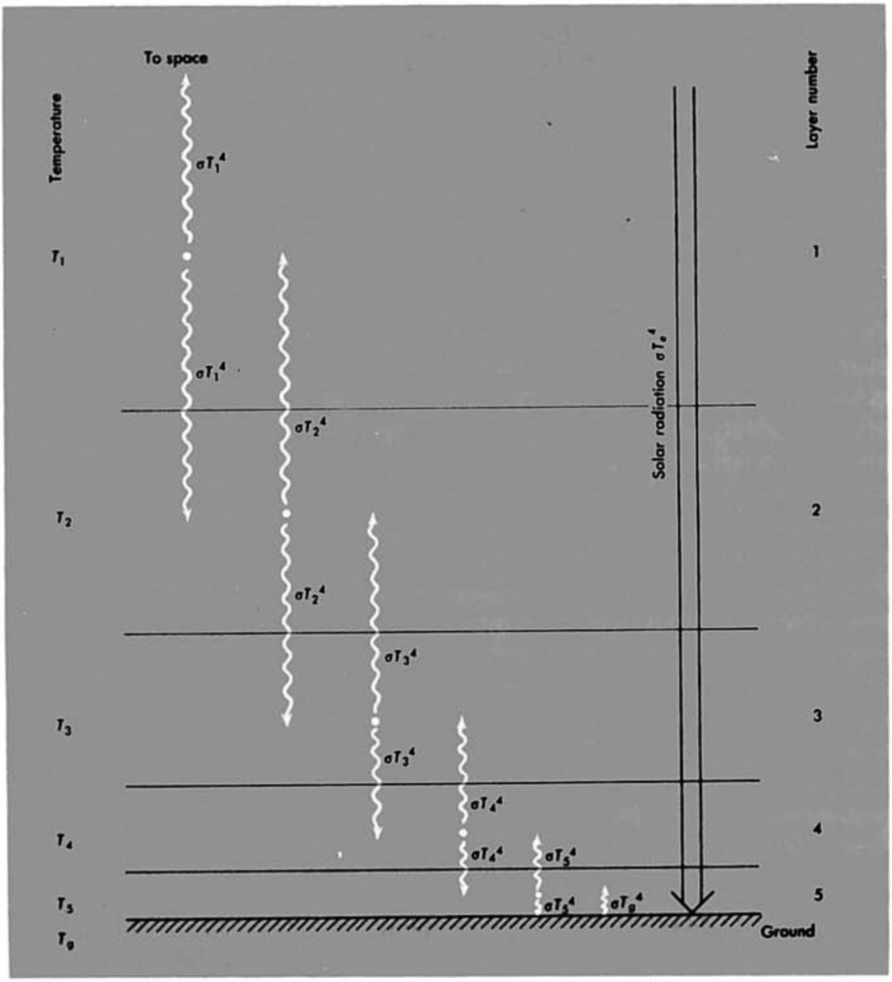

FIGURE 3-6 *Layers of the atmosphere exchanging radiant heat energy with adjacent layers. For this illustration we have chosen an atmosphere that is transparent to solar radiation, but which has an optical thickness of five for planetary radiation.*

In the same way, if we equate the energy radiated by the second layer from the top to the energy absorbed we find

$$\sigma T_2^4 \text{ (upward)} + \sigma T_2^4 \text{ (downward)} = \sigma T_3^4 \text{ (upward)} + \sigma T_1^4 \text{ (downward)} \quad (3\text{–}7)$$

since this layer is heated by layer 1 above as well as by layer 3 below. We find

$$T_3{}^4 = 2\,T_2{}^4 - T_1{}^4 = 3\,T_1{}^4 = \;3\,T_e{}^4 \qquad (3\text{–}8)$$

where we have used the fact that $T_2{}^4$ equals $2T_e{}^4$ (Eq. 3–6).

From the energy balance in each successive layer, working down from the top of the atmosphere, we can deduce the temperature of each layer. We have carried the process far enough to see what the answers must be: $T_4{}^4$ is $4T_e{}^4$, $T_5{}^4$ is $5T_e{}^4$, and so on.

Finally, consider what happens at the planet's surface. The energy absorbed is $\sigma T_e{}^4$ from the sun, plus $\sigma T_5{}^4$ $(=5\sigma T_e{}^4)$ from the atmosphere. Energy lost is $\sigma T_g{}^4$ where T_g is the ground temperature. Hence (canceling σ's)

$$T_g{}^4 = T_e{}^4 + 5\,T_e{}^4 = 6\,T_e{}^4 \qquad (3\text{–}9)$$

It is simple to extend the result to an atmosphere of arbitrary optical thickness (and we shall need the result later). We find

$$T_g{}^4 = (1 + \text{optical thickness}) \times T_e{}^4 \qquad (3\text{–}10)$$

Our result is illustrated in Fig. 3–7, where we have plotted, against height, the fourth power of the temperature divided by the temperature in the top layer of the atmosphere, T_1. The figure shows that the temperature decreases steadily with height but approaches a constant value at high altitudes.

In order to plot Fig. 3–7 we have to choose the layers so that each appears equally opaque to planetary radiation. The thickness of the layers depends, in general, on the gas concerned, the optical properties of the gas, and the way in which the gas is distributed in the vertical. To illustrate, let us assume that the absorbing gas is the only gas in the atmosphere and that the density decreases exponentially with a scale height, H.

We have to divide the atmosphere into five slabs, with the same amount of gas in each slab. According to the discussion in Chapter 1, the change in pressure between the top and bottom of a slab is equal to the mass of gas per unit area in the slab times the gravitational acceleration. Thus, the slabs are bounded by surfaces at which the pressure is equal to $\frac{4}{5}$, $\frac{3}{5}$, $\frac{2}{5}$, and $\frac{1}{5}$ times the pressure at ground level.

We have computed the average temperature of each of these layers. The temperatures should now be plotted at the heights corresponding to the middle of each layer, that is, at the levels where the pressure is $9p(0)/10$, $7p(0)/10$, $5p(0)/10$, $3p(0)/10$, and $p(0)/10$, where $p(0)$ is the pressure at the ground.

Now consider the barometric law in the form given in the footnote on p. 9. We can write

$$\frac{z}{H} = -\left\{\frac{1}{\log e}\right\} \times \left\{\log \frac{p(0)}{p(z)}\right\} \qquad (3\text{–}11)$$

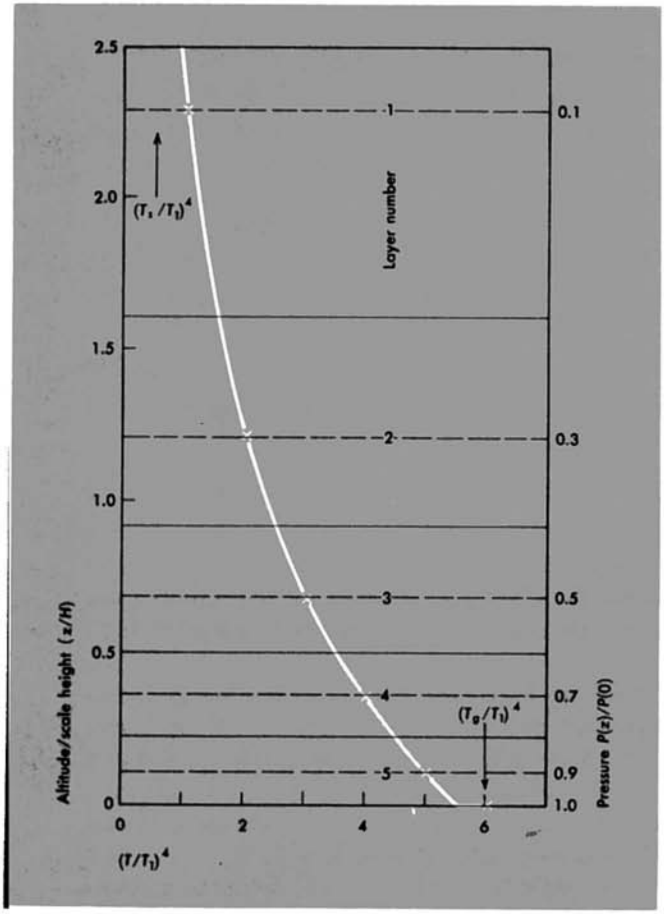

FIGURE 3-7 *The variation of temperature with altitude in an atmosphere of optical thickness 5. The crosses are the results of the calculations in this chapter. The curve is the result of a more detailed calculation. The horizontal solid lines mark the boundaries of the five layers. The broken lines show the mid-points; one-half of the mass of each layer lies on each side of a broken line. The quantities T_1, T_s, and T_g are explained in the text.*

The temperatures we have calculated should be plotted at values of z/H corresponding to $p(z)/p(0)$ equal to $9/10$, $7/10$, $5/10$, $3/10$, and $1/10$. Since $\log e = 0.434$, we find, using logarithms, the values 0.11, 0.36, 0.69, 1.21, and 2.3 for z/H. A similar calculation gives the positions of the surfaces that separate the layers themselves. The layers and midpoints calculated in this way are used in Fig. 3–7. If we are interested in actual heights, then we must multiply z/H by the scale height. For example, if $H = 8.4$ km, as on Earth, the lowest broken line is at a height of 0.92 km and the uppermost is at a height of 19.4 km.

We have now developed the theory of radiative transfer to the point where we can examine, in a quantitative manner, the greenhouse effect on a real planet. An interesting planet in this regard is Venus because Venus has a ground temperature of about 700°K but has an effective temperature of only 243°K. Can the high ground temperature on Venus be a result of the greenhouse effect?

The Greenhouse Effect on Venus

Let us ask how great an optical thickness would be required for the atmosphere of Venus if the greenhouse effect were to provide the high ground temperature. Remember that the optical thickness is the total number of layers into which the atmosphere must be divided so that radiation emitted by each layer is absorbed in the immediately adjacent layers. A massive atmosphere composed of gases that absorb strongly in the infrared region of the spectrum has a large optical thickness. A thin atmosphere has a relatively small optical thickness.

From Eq. (3–10) we can calculate the optical thickness for Venus if we know the effective temperature (243°K) and the ground temperature (700°K) and if we assume that the atmosphere is in radiative equilibrium. We find

$$\text{Optical thickness} = \left\{\frac{700}{243}\right\}^4 - 1 = 68 \qquad (3\text{–}12)$$

Does the atmosphere of Venus have such a large optical depth, and if it does, is the greenhouse theory the explanation of the high surface temperature?

The first question can be answered to some degree. Carbon dioxide alone cannot give such a large optical depth. The problem lies in the uneven efficiency of absorption in the infrared, which occurs principally in narrow regions of the spectrum (see Fig. 3–4). We require a large optical depth throughout the spectrum including the gaps where absorption is weak. The amount of carbon dioxide known to exist on Venus cannot give the required optical thickness. It is possible that the more transparent gaps are blocked by absorption by other gases that happen to absorb strongly where carbon dioxide absorbs weakly. Water vapor has been proposed as such a supplemental absorber, although we have no direct evidence for its existence in sufficient quantities on Venus. Alternatively, it is possible that the clouds themselves may perform a similar function absorbing strongly throughout the infrared spectrum.

This last suggestion, like all others, has difficulties, and we will not pursue the Venus greenhouse effect further at this stage. We introduced it solely to illustrate principles. It is, after all, quite certain that our model is too simple to explain every feature of the Venus surface temperatures. For example, a radiative equilibrium model predicts on Venus, as on Earth and Mars, high temperatures in the tropics, where sunlight is most intense, and low temperatures in the polar regions. In fact, no significant temperature difference has yet been observed between these locations on Venus. Further discussion must be postponed to Chapter 4, where we examine the effect of winds on atmospheric temperatures.

The Greenhouse Effect on Earth

Let us make a second application of the radiative transfer theory that we have derived, this time to a calculation of the profile of temperature as a function of altitude in the atmosphere of the Earth. Although we know much more about Earth's atmosphere than we know about the atmosphere of Venus, it is not entirely clear how we should divide the atmosphere into layers when we have transparent and opaque spectral regions, as illustrated in Fig. 3–4.

Experience suggests that a reasonable model consists of two layers, with the top layer centered at a height of about 3 km and the bottom layer centered at a height of about 0.5 km. We can derive these heights by the method outlined earlier, if we remember that water vapor is the principal absorbing gas in the Earth's atmosphere. Observations show that the scale height of water vapor is about 2 km.

The temperature of the top layer is equal to the effective temperature. For Earth, this is 253°K (shown by an X in Fig. 3–8 at a height of 3 km). For the bottom layer, the fourth power of the temperature is equal to twice the fourth power of the effective temperature because the bottom layer is the second layer from the top. We find that the temperature of the bottom layer is 297°K (shown by an X in Fig. 3–8 at a height of 0.5 km).

Since these two data are not enough to draw a profile, we will compute one further point, namely, the lower limit to the temperature at very great heights. We hinted at the existence of such a limit in Fig. 3–7. A detailed computation shows that the temperature never drops below $(T/T_e)^4 = 1/2$ and tends to this limit as the height increases.

To show that this is the appropriate limit, consider Fig. 3–9, which shows a very thin layer high in the atmosphere, and hence, far above the average height of layer 1. The *opacity* of this layer is ε and this we assume to be a very small quantity.

The opacity specifies the capacity of a layer to absorb radiation. Hence, from the upwelling radiation from layer 1, which equals σT_1^4, a fraction $\varepsilon\sigma T_1^4$ is absorbed in the thin layer. This is indicated in Fig. 3–9 by a beam of flux σT_1^4 incident from below and a beam of flux $\sigma T_1^4(1-\varepsilon)$ leaving to space.

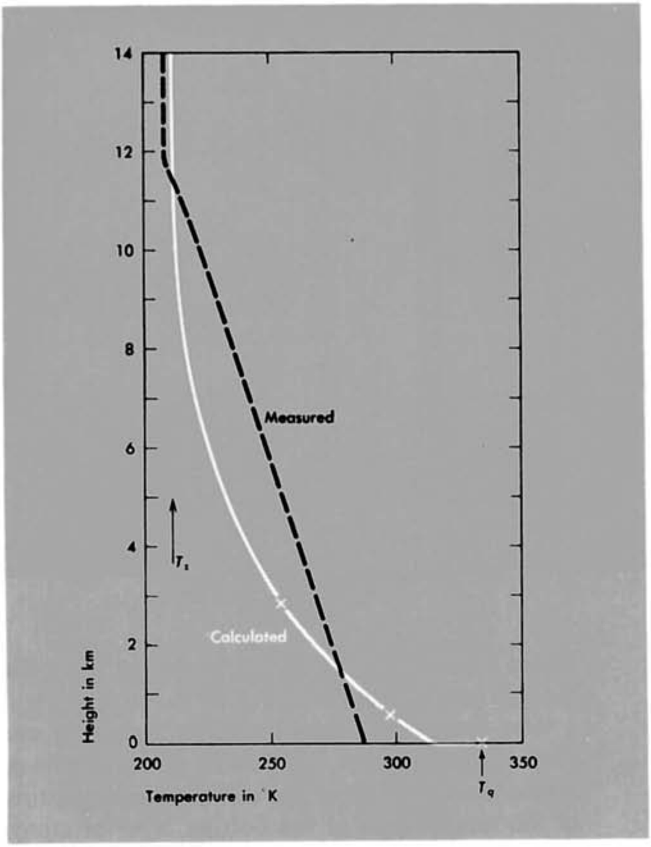

FIGURE 3-8 *The temperature profile in the Earth's atmosphere calculated under the assumption of radiative equilibrium compared with average measured temperatures.*

The temperature of the thin layer is T_s, called the *skin temperature.* If the layer were opaque, it would emit σT_s^4 in both directions, up and down. It is not opaque, however, because it is very thin, and we must appeal to one of *Kirchoff's radiation laws*, which tells us that for translucent bodies *emissivity* and *opacity* are equal; both are therefore equal to ε. The total emission by the layer is therefore $2\varepsilon\sigma T_s^4$, one half upward and one half downward.

If the layer is in radiative equilibrium, energy absorbed equals energy emitted or

$$\varepsilon\sigma T_1^4 = 2\varepsilon\sigma T_s^4 \tag{3–13}$$

Upon rearranging, we find

$$\left(\frac{T_s}{T_1}\right)^4 = \frac{1}{2} \tag{3–14}$$

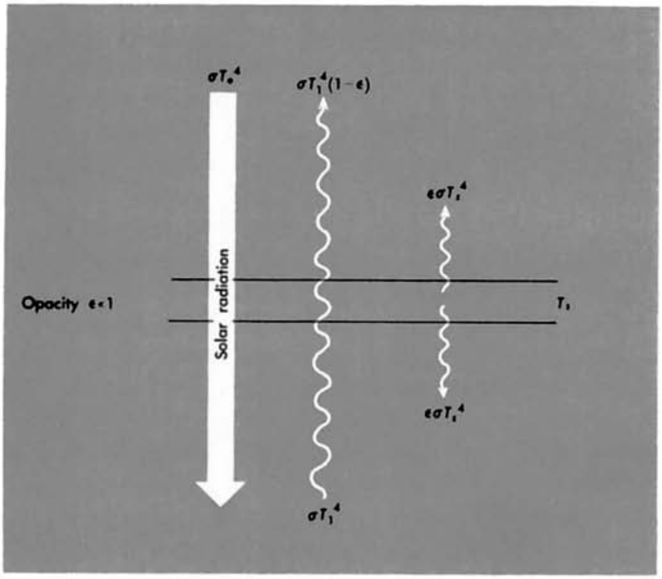

FIGURE 3-9 *Radiative equilibrium in a very thin layer at the outer limit of the atmosphere.*

which is the result we set out to derive.* For Earth, with $T_1 = T_e = 253°K$, the skin temperature is $T_s = 212°K$.

Our theoretical model will be complete once we have calculated the ground temperature. The fourth power of the ground temperature is equal to the fourth power of the effective temperature added to the fourth power of the temperature of the bottom layer of atmosphere. We find a value for the ground temperature of 333°K, shown by an arrow at the bottom of Fig. 3–8.

We can now compare the theoretical temperatures we have just calculated with average temperatures measured in the real atmosphere. Measured temperatures are shown by the broken line in Fig. 3–8. We see that the theoretical model is quite successful at altitudes above 10 km but that there are substantial deviations throughout the troposphere. Our theory is inadequate because radiation is not the only process that carries heat upward from the ground and from the lower levels of the troposphere. Another process tending to hold down the temperature at the ground and to increase the temperature of the upper troposphere is known as *convection*. Under certain circumstances convection can carry heat away from the ground more effectively than can radiation.

*Higher temperatures result if the upper levels of an atmosphere absorb solar radiation directly, as is the case in the terrestrial ozone layer. This topic is developed in the Appendix.

Heat Transport by Convection

The mechanism of convection is familiar to all of us. It occurs because air expands when it is heated, causing the density to decrease (see our discussion of the ideal gas law in Chapter 1). If we have abnormally hot air near the ground, it will be lighter than its surroundings and, like a balloon, will attempt to rise. Similarly, if there is abnormally cool air at high levels, it may be dense compared with its surroundings and may tend to sink. Low level warm air will replace high level cold air and vice versa, giving a net upward transport of heat. We must now determine the conditions under which this process of convection takes place and how it affects the temperature structure of an atmosphere.

Our first problem is that the temperature of a rising parcel of air tends to fall as the parcel rises because the atmospheric pressure decreases with altitude at a rate given by the barometric law (see Chapter 1). Thus, a rising parcel of air moves from a region of higher pressure to a region of lower pressure and expands as it does so. The temperature of the air decreases as the air expands, just as does the temperature of air released from a tire. The expanding parcel pushes back the air pressing on its boundaries and, therefore, does work upon the surroundings. Work is a form of energy, and conservation of energy requires that some other form of energy in the parcel must decrease. The only possibility for dry air is the internal thermal energy of the molecules. Consequently the temperature must drop as the parcel rises.

To an acceptable degree of approximation we may regard this decrease of temperature with height as independent of the state of the surrounding atmosphere. In Fig. 3–10, the solid line *AB* represents the variation of temperature with height in a parcel of air rising through the atmosphere. The rate of decrease of temperature with height corresponding to *AB* is called the *adiabatic lapse rate* (deg cm^{-1}). We shall calculate its value below. Let us accept its existence for now and consider what happens in an atmosphere when temperature varies at a different rate. If the temperature of our rising parcel of air decreases more rapidly than the temperature of the surrounding atmosphere, the parcel will always be colder and denser than the atmosphere around it (see Fig. 3–10). Under these circumstances its excess weight will tend to drag it back down to the atmospheric level from whence it came, and convection will not occur. The atmosphere is then said to be *stable.*

On the other hand, if the temperature of the atmosphere decreases with altitude more rapidly than the temperature in a rising parcel of air, we

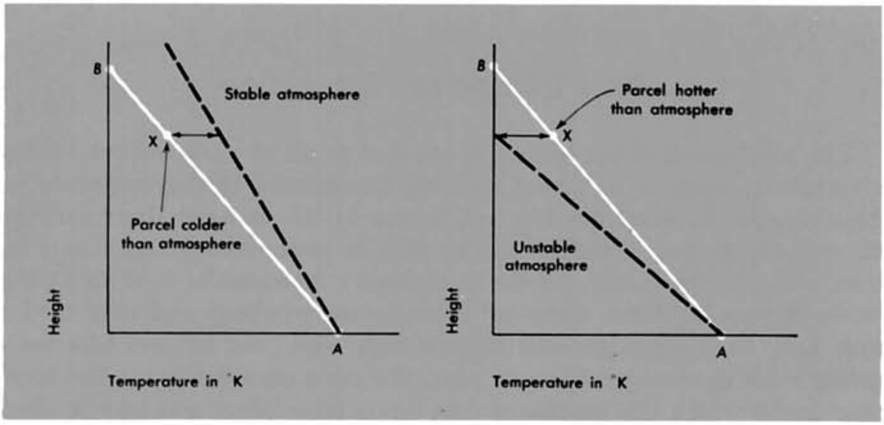

FIGURE 3-10 *The temperature profile in a stable atmosphere is shown by the dashed line in the left-hand diagram. The dashed line in the right-hand diagram shows the temperature profile in an unstable atmosphere. The solid line in each diagram has a negative slope equal to the adiabatic lapse rate; it is approximately the temperature of a parcel of air rising rapidly through the atmosphere. The parcel starts at the bottom point A in each case, where it has the same temperature as the surrounding atmosphere and rises along the path AB. When the parcel on the left reaches point X, it is colder than the surrounding atmosphere and therefore denser. It tends to sink back to the level where it started. When the parcel on the right reaches point X, it is hotter and therefore less dense than the surrounding atmosphere. Buoyancy causes it to rise still further.*

have the situation we envisaged when we began the discussion of convection. After having risen a short distance, the parcel of air will have a higher temperature than the surrounding atmosphere in spite of its temperature having fallen as a result of expansion. Because it is hotter, the density of the air in the parcel will be lower than the density of the atmosphere, and buoyant forces will accelerate its upward motion. Thus, the atmosphere is in a condition whereby it can change spontaneously, without outside intervention, like a pen balanced precariously on its tip. This circumstance we term ***unstable.***

We conclude that convection can only occur spontaneously if the temperature of the atmosphere decreases quickly enough with altitude. An isothermal atmosphere, for example, is stable against convection. Although the temperature in the troposphere generally decreases with increasing altitude, there are times when the air near the ground is colder than the air above, giving a temperature that increases with altitude instead. This meteorological phenomenon is called an ***inversion.*** During an inversion, convection is strongly inhibited and very little mixing occurs from one level of the atmosphere to another. As a result, pollutants accumulate in the air near the ground instead of being dispersed to higher levels. It is at times of inversion that air pollution is most severe.

We have already pointed out that rising parcels of air follow temperature curves very similar to each other (curve *AB* in Fig. 3–10). We can evaluate the slope of this line relatively easily when the surrounding atmosphere also has temperatures lying on the curve *AB*: a parcel of air then can rise and fall in a neutral fashion without exchanging heat with the surroundings. When no heat is exchanged, we refer to the change as *adiabatic.* The calculation in the Appendix shows that the temperature falls at a rate

$$\text{Adiabatic lapse rate (deg cm}^{-1}) = \frac{\text{Acceleration of gravity (cm sec}^{-2})}{\text{Specific heat of air at constant pressure (erg deg}^{-1}\text{ gm}^{-1})} \quad (3\text{–}15)$$

Values of the adiabatic lapse rate of different planets are given in Table 3–2.

Convection is an efficient means of transporting heat in a gas. Both experience and theory show that the rate of decrease of temperature with height in a planetary atmosphere cannot exceed the adiabatic lapse rate by a substantial amount except very close to the ground, where convection is inhibited. Imagine an atmosphere in which the temperature lapse rate is initially less than the adiabatic lapse rate and imagine that the lower levels are heated while the upper levels are cooled, so that the lapse rate increases with time. Initially the atmosphere is stable against convection and no moion occurs. As soon as the lapse rate exceeds the adiabatic lapse rate, however, the atmosphere becomes unstable against convection. Convection carries heat from the hot lower levels of the atmosphere to the cold upper levels, inhibiting further increase of the lapse rate.

We can now decide whether or not the theoretical profile of temperature in a planetary atmosphere is stable against convection. When the predicted rate of decrease of temperature with altitude is less than the adiabatic lapse rate, the atmosphere is stable, and convection will not occur to modify the temperature profile. But whenever the predicted lapse rate

Table 3–2

Adiabatic Lapse Rates

Planet	Gas	Gravitational Acceleration (cm sec^{-2})	Specific Heat (erg gm^{-1} deg^{-1})	Adiabatic Lapse Rate (deg km^{-1})
Venus	CO_2	888	8.3×10^6	10.7
Earth	N_2,O_2	981	1.0×10^7	9.8
Mars	CO_2	373	8.3×10^6	4.5
Jupiter	H_2	2620	1.3×10^8	20.2

exceeds the adiabatic lapse rate, convection will occur, and the temperature profile will be modified so that temperature decreases at about the adiabatic rate.

Our examination of the temperature profile in an atmosphere transporting heat only by radiation showed that the lapse rate is small at high altitudes, where there is little absorbing gas, but increases steadily as the altitude decreases; close to the ground it becomes very steep indeed, as indicated by the horizontal portions of Fig. 3–7 and Fig. 3–8. We can, therefore, conclude that an atmosphere will be stable against convection at higher altitudes but may become unstable at the lower altitudes where there is enough gas to absorb infrared radiation strongly.

These considerations and their effect on the temperature profile for Earth, are shown in Fig. 3–8. At altitudes above 12 km, the temperature is not influenced by convection, and the measured temperature profile follows the radiative equilibrium solution. At lower altitudes, convection takes over from radiation as the most important heat transport process and the temperature profile becomes a straight line.* The net effect of convection, as the figure shows, is to reduce the ground temperature by about 60°K.

The theory we have described, including heat transport by convection as well as by infrared radiation, is reasonably successful at explaining the average temperature profile in the lower few tens of kilometers of the Earth's atmosphere. The layer at the bottom of Fig. 3–8, where convection is important and the temperature decreases steadily with increasing height, corresponds to the troposphere shown in Fig. 3–1. The overlying layer, where convection is not important and the temperature is very nearly constant, corresponds to the lower stratosphere. The temperature profile in the upper stratosphere and mesophere can also be understood in terms of radiative transfer and convection but, as we have already noted, it is necessary to take into account the heat source provided, at heights around 50 km, by the absorption of solar ultraviolet radiation by ozone (see Appendix).

The radiative-convective theory is not applicable at heights above about 80 km in the Earth's atmosphere where the air is too thin to interact strongly with radiation. We shall discuss this high altitude region (the thermosphere) later in this chapter. Before we do so, let us consider the use of the radiative-convective theory to predict temperatures in the lower atmospheres of some of the other planets.

First consider Venus. If we apply the radiative-convective theory, we find that beneath the cloud tops the temperature should vary at the adiabatic lapse rate. As far as we can tell from space probe measurements,

*The average lapse rate in the troposphere is 6.5°K km^{-1}, which is smaller than the adiabatic lapse rate given in Table 3–2. The reasons for this difference are known. We have correctly described some essential physical processes, but not all of them. Two other factors are large-scale planetary motions (discussed in Chapter 4) and the condensation of water (discussed in Chapter 5).

this prediction is verified. Introduction of convection, however, does not eliminate the difficulty that the temperature is the same at the equator as at the poles. Our theory still predicts that σT_e^4 is equal to the absorbed flux of solar radiation; although this is small at the poles and large at the equator, no variation of T_e is, in fact, observed. Thus, while these ideas are helpful in picturing factors involved in the high ground temperatures and the adiabatic lower atmosphere, they are definitely over-simplified.

The Martian Troposphere

For Mars, let us first consider an average over day and night conditions in the Martian tropics. The theory predicts an average temperature profile that resembles the terrestrial profile shown in Fig. 3–8, except that temperatures are lower on Mars because Mars is farther from the Sun. In the bottom 15 km of the tropical Martian atmosphere the temperature decreases at the adiabatic lapse rate from a value of about 230°K near the ground. Above 15 km the temperature is nearly constant and has a value of about 155°K. The region of temperature decrease corresponds to the troposphere on Earth; convection plays a dominant role in maintaining the average temperature profile. The overlying region corresponds to the lower stratosphere on Earth, where heat is transported mainly by radiation. We have already mentioned that we do not expect to find a temperature maximum at intermediate heights in the Martian atmosphere, corresponding to the terrestrial stratopause (Fig. 3–1), because Mars does not have enough oxygen in its atmosphere to produce ozone.

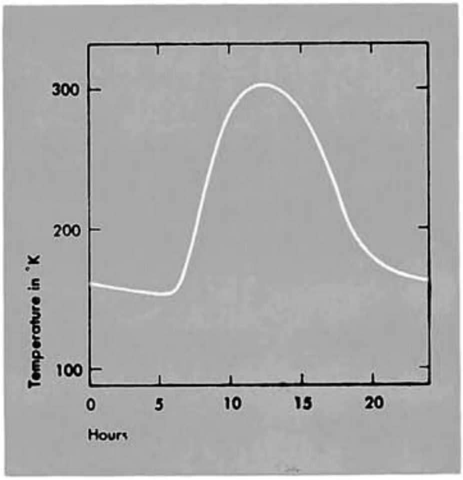

FIGURE 3-11 *Diurnal variation of the ground temperature at the equator of Mars during an equinox. The temperature during the day has been measured. The night-time values are theoretical. (After P. Gierasch and R. M. Goody, 1968.)*

The Martian troposphere differs from the terrestrial troposphere in one important respect. On Mars there is a large diurnal variation of the temperature. The temperature of the ground at the equator of Mars is shown in Fig. 3–11. At noon it is about 300°K, not very different from temperatures in the Earth's tropics. At night, on the other hand, the Martian temperature drops to a frigid 160°K, much colder than any place on the surface of the Earth. This large diurnal oscillation is in part a result of the absence of oceans on Mars. Our oceans are partially transparent to sunlight and are also in a continual state of motion. Thus, a substantial layer at the top of the ocean absorbs heat during the day and loses heat during the night. A thick layer of water can hold a large amount of heat, so very little temperature change results from the diurnal cycle of heating and cooling.

On Mars, however, in the absence of oceans, all the daytime heating and night-time cooling must take place right at the surface of the ground. The thermal conductivity of the Martian surface is believed to be small, with only a thin layer taking part in the diurnal oscillation. Since a thin layer of dry surface material can hold very little heat, large temperature changes result from the diurnal cycle of heating and cooling. (We can observe this phenomenon on Earth, where diurnal surface temperature changes are much larger in dry desert areas a long way from the sea than they are on or near the oceans or on land covered with vegetation. Observations on Mars indicate that the surface does indeed resemble light, dry desert sand.)

The large variation in the ground temperature on Mars produces a correspondingly large diurnal variation in the flux of infrared radiation. Since this flux is mainly responsible for heating the lower atmosphere, we may expect a corresponding variation in atmospheric temperature. Because of its composition, the Martian atmosphere is much more sensitive than the terrestrial atmosphere to changes in infrared flux.

On Mars the atmosphere contains a large proportion of carbon dioxide molecules (see Chapter 1), which absorb infrared radiation. On Earth, on the other hand, the most important absorbers are water vapor molecules, which constitute a very small fraction of the total atmosphere. Heat absorbed by a molecule of water vapor on Earth must be shared amongst hundreds or thousands of oxygen and nitrogen molecules which themselves do not absorb heat radiation. On Mars, however, each carbon dioxide molecule absorbing or emitting thermal radiation can use the heat to change its own thermal energy. The response of the atmospheric temperature to changes in radiation is, therefore, far more rapid for Mars than for Earth; so the large diurnal temperature oscillation on Mars occurs not only at the ground but is also transmitted by radiation into the atmosphere up to a height of several kilometers. Theoretical profiles of Martian temperatures at different times of day are shown in Fig. 3–12. On Earth, by way of con-

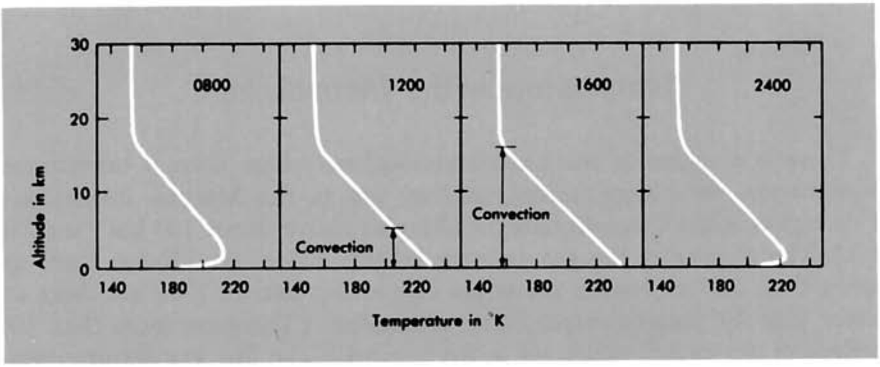

FIGURE 3-12 *Theoretical temperature profiles in the Martian atmosphere at different times of the day (after P. Gierasch and R. M. Goody, 1968.) At 0800 hours the ground is colder than the overlying atmosphere, and there is no convective region. Convection starts before 1200 hours, is stronger at 1600 hours, but dies away at night. At 2400 hours the ground is colder than the atmosphere once again. Ground temperatures are given in Fig. 3-11. There is a shallow layer near the ground in which temperature changes very rapidly with height.*

trast, diurnal temperature oscillations are small except in the few tens of meters closest to the ground.

The diurnal variation of temperature in the lowest levels of Mars' atmosphere causes a corresponding variation in the occurrence of convection. During much of the night, as Fig. 3–12 shows, the ground is colder than the lower atmosphere, and the temperature actually increases with height in the first one or two kilometers (there is an inversion). A temperature profile such as this is stable against convection. Even at higher altitudes, where the temperature does decrease with height, it does so at a rate less than the adiabatic lapse rate and the atmosphere remains stable. The result is that there may be no region of convective instability in the Martian atmosphere at night.

After the sun rises, however, theory indicates that the ground temperature increases rapidly to values considerably in excess of the atmospheric temperature. A region of convective instability develops, close to the ground at first, but extending higher and higher as the atmosphere warms up. By the end of the day Mars has a well-developed convective layer extending as high as 15 km. Soon after nightfall the ground and the lower atmosphere cool down again, and convection ceases. The predicted diurnal variation in the thickness of the convective layer is, as far as we know, peculiar to Mars.

Temperature in the Thermosphere

There is a region of the Earth's atmosphere where diurnal temperature changes are very large indeed, as they are in the Martian troposphere. This region is the thermosphere, at altitudes above about 100 km (see Fig. 3–1). The density of the gas at these heights is low. At 120 km there are fewer than 10^{12} molecules per cubic centimeter, and at 1000 km there are fewer than 10^{6} molecules per cubic centimeter. (There are more than 10^{19} molecules per cubic centimeter at the ground.) The low gas density means that the thermosphere can hold very little heat, and this means that the temperature of the thermosphere can respond rapidly to changes in the amount of solar radiation absorbed at these levels. The result is a substantial diurnal variation in thermospheric temperature, with afternoon temperatures exceeding early morning temperatures by 300°K or more.

The gas densities in the thermosphere are low, but the temperatures are generally high. The mean temperature of the neutral gas increases from 350°K at 120 km to about 1000°K at 250 km. Above this height the temperature is very nearly constant. One might think that these high temperatures are evidence of a large source of energy, but such is not the case. The total rate of heating in the thermosphere is only about one-millionth of the rate at which the ground is heated by visible radiation from the sun. Instead, the high temperatures reflect the inefficiency of mechanisms for removing heat from this region of the atmosphere. For although in the lower atmosphere heat is transported efficiently by infrared radiation emitted by water vapor, carbon dioxide, and ozone, in the thermosphere infrared radiation is relatively unimportant.

Temperature, as normally defined, has to do with the energy of translational motion of the molecules. Absorption and emission of infrared radiation, on the other hand, are concerned with the increase or decrease of the rotational and vibrational energy of molecules. If collisions are rapid, these three types of energy—translation, rotation, and vibration—readily interchange. An excess translational energy leads to excess vibrational energy that can then be emitted as heat radiation. In this manner heat radiation affects the temperature of the gas as described by the Stefan-Boltzmann law.

If the pressure is very low, however, and collisions are infrequent, the different forms of energy are not rapidly shuffled. High temperature leads to high velocities and high translational energy but not necessarily to the high vibrational and rotational energies that affect heat radiation. This problem is illustrated schematically in Fig. 3–13. Up until now we have assumed the energy transfer link (bottom of the diagram) to be rapid; thus,

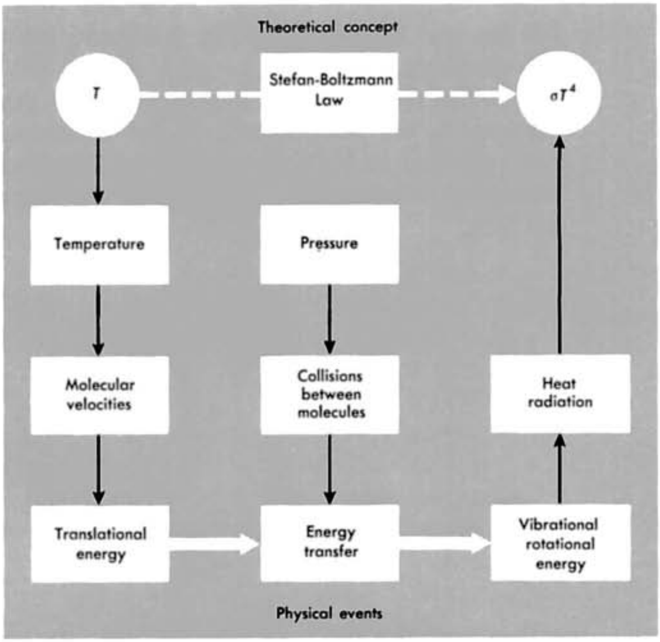

FIGURE 3-13 *The relationship between the theoretical concept of heat radiation (Stefan-Boltzmann law) and the physical chain of events. If pressure is low, the rate of energy transfer is low and high temperature no longer implies high heat radiation or vice versa.*

temperature T led naturally to heat radiation σT^4. But if the link is slow because of infrequent collisions, heat radiation has little connection with temperature. This is the situation in the thermosphere. If we want to lose or gain heat in order to change the temperature, we must find another way besides thermal radiation.

The other form of heat transfer that we have discussed so far is convection. The thermosphere, however, is strongly stable against convection because temperature increases with height, and we anticipate almost no heat transfer from this cause. The only remaining possibility is the transfer of heat by direct collisions between molecules from high temperature regions and molecules from low temperature regions. This is the mechanism of thermal conduction. Since conduction in a gas is a relatively inefficient means of transporting heat, large temperature differences are needed to conduct away the small amount of solar heat that is absorbed in the thermosphere. The small thermospheric heat source is provided by solar ultraviolet radiation with wavelengths shorter than 1000 Å, which is absorbed at heights of about 200 km (see Fig. 2-4). This is the radiation that photoionizes the atmosphere and produces the ionosphere, a subject we discussed in Chapter 2.

The heat must be conducted all the way down to about 100 km, where collisions are sufficiently frequent to allow infrared emission to remove the heat. This is the reason for the substantial increase in temperature between

100 km and 200 km (see Fig. 3–1) and for the high temperature in the thermosphere.

Satellite measurements over a period of years have revealed wide fluctuations in the temperature of the thermosphere; daytime temperatures have climbed as high as 1800°K and have fallen as low as 900°K. These excursions are caused by changes in the flux of ultraviolet radiation from the Sun.