12.1 Planetary Radiation Balance

The amount of solar radiation received by the Earth and its atmosphere is equal to the solar constant minus the amount of shortwave radiation reflected to space, times the cross-sectional area of the Earth that is perpendicular to the beam of parallel solar radiation. Assuming that the terrestrial emission is equivalent to the blackbody flux at temperature T_e, the amount of longwave radiation emitted by the Earth is the equivalent blackbody flux times the surface area of the Earth. Hence we can write the following expression for the Earth's radiative energy balance under conditions of radiative equilibrium

$$S\left(1-\alpha_p\right)\pi r^2 = \sigma T_e^4\, 4\,\pi r \tag{12.1a}$$

or

$$\frac{S}{4}\left(1-\alpha_p\right) = \sigma T_e^4 \tag{12.1b}$$

where r is the radius of the solid Earth and α_p is the planetary albedo. The factor 4 in the denominator arises from the ratio of the surface area of a sphere to its cross-sectional area. The amount of solar radiation intercepted by the Earth is $S\pi r^2$. The globally averaged insolation at the top of the atmosphere is $S\pi r^2/4\pi r^2 = S/4 = 342$ W m^{-2}.

Solving (12.1b) for the equivalent black body temperature, T_e, we obtain

$$T_e = \sqrt[4]{\frac{S\left(1-\alpha_p\right)}{4\sigma}} \tag{12.1c}$$

The temperature T_e is not necessarily the actual surface or atmospheric temperature of the planet; it is simply the equivalent blackbody emission temperature a planet requires to balance the solar radiation that it absorbs. Using a value of $\alpha_p = 0.31$, we obtain from (12.1c) a value of $T_e = 254$ K. Note that this temperature is much less than the observed global mean surface temperature $T_0 = 288$ K. The difference between T_e and T_0 arises from the emission of thermal radiation by atmospheric gases and clouds at temperatures colder than T_0.

14.3 Planetary Energy Balance

The planetary energy budget for Earth was described in Section 12.1. There are five primary physical characteristics that control the radiation balance of a planet:

(i) The distance of the planet from the sun determines solar illumination.

(ii) The mass and composition of the planet. The mass determines the planetary escape velocity while the composition of the core determines the presence and strength of the planetary magnetic field.
(iii) The geological structure of the planet determines the outgassing of planetary effluents into the atmosphere.
(iv) Orbital parameters determine day and year length.
(v) Internal planetary heat sources.

Secondary controls on planetary energy budgets arise from:
(v) The evolving planetary albedo.
(vi) The infrared radiative properties of the planetary atmosphere.

14.3.1 The Planetary Greenhouse Effect

In Section 12.1, the radiative balance for a planet in radiation equilibrium was discussed, whereby the outgoing longwave radiation of the planet is equal to the net incoming solar radiation. The inner planets are essentially in radiation equlibrium. Jupiter, Saturn, Neptune, and possibly Uranus, however, are not in radiative equlibrium. Each of these planets actually radiaties more energy to space that it receives from the Sun: Jupiter, by a factor of 1.9; Saturn, by a factor of 2.2, and Neptune by a factor of 2.7. The generally close similiarity in physical properties between Uranus and Neptune suggests that if Neptune has an internal heat source, then Uranus ought to also. The enhanced infrared emissions from the Jovian planets are apparently a product of the planet's own internal heat source. It has been theorized that these planets are evolutionary forms of a low mass star. Hence the present day radiation energy may come from an internal reservoir of thermal energy generated by slow shrinkage of the entire planet..

For the inner planets, which are in planetary radiative equilibrium with the Sun, the planetary radiation balance is written following (12.1) as

$$T_e^* = \left(\frac{S(1-\alpha_p)}{4\sigma}\right)^{1/4}$$

The temperature, T_e*, is defined as equivalent black body temperature at which the planet and its atmosphere radiates to space. Table 14.6 lists the equivalent black body temperatures for the inner planets, along with values of the solar constant, S, and planetary albedo, α_p. Values of T_e* decrease generally with distance of the planet from the sun, with the modulation of planetary albedo by clouds also providing an important control on T_e*.

Table 14.6: Values of equivalent black body planetary temperature, T_e*, as a function of the planetary solar constant, S, and planetary albedo, α_p, and optical thickness of the atmosphere n.

Planet	α_p (%)	S (W m^{-2})	T_e* (K)	T_0 K	n
Mercury	5.8	9200	442	452	0.09
Venus	71	2600	244	740	84
Earth	33	1370	254	288	0.65
Mars	17	600	217	223	0.12

For a planet in radiative equilibrium, comparison of T_e* with T_0 gives an indication of the planetary greenhouse effect. Note the especially large greenhouse effect of Venus. Mercury and Mars, with their tenuous atmospheres, have little greenhouse effect. Earth has a significant greenhouse effect, but nothing approaching that of Venus.

To examine the planetary greenhouse effect, consider a simple atmosphere characterized as follows (see Figure 14.8):

(i) The atmosphere is made up of n slabs, each with temperature T_j .

(ii) Each slab contains a gas that absorbs longwave radiation. The thickness of each slab is such that it is just optically black with an emissivity of unity. That is, all longwave radiation entering the slab will just be absorbed.

(iii) The longwave absorbing gas is equally mixed by mass throughout the atmospheric column. Thus, in order for each slab to be just optically black, it must contain the same mass of absorbing gas.

(iv) The slabs are assumed to be transparent to visible radiation. In this simplified atmosphere, only the surface reflects solar radiation.

(v) All other heating or transport processes are ignored.

Considering only radiative processes and assuming equilibrium, the energy balance in slab j is given by

$$2F_j = F_{j-1} + F_{j+1} \tag{14.10}$$

At the surface, the energy balance is

$$S\left(1 - \alpha_0\right) + F_n = \sigma T_0^4 \tag{14.11}$$

where the n^{th} layer is the layer just above the surface, and the surface is assumed to emit as a black body. Since each of the layers is assumed to be black, we can use the Stephan-Boltzmann law to write $F_j = \sigma T_j^4$ and so on. Since the effective temperature is the radiating temperature of the uppermost black layer of the atmosphere, we can also write

$$F_1 = \sigma T_e^{*4} \tag{14.12}$$

By combining (14.10)-(14.12) and examining Figure 14.8, we can derive a relationship between the radiative surface temperature of a planet (T_0) and the effective temperature

$$T_0 = (1 + n)\, T_e^{*} \tag{14.13}$$

where n is the number of optically black layers in the column or the optical depth.

From observations of T_0 and calculations of T_e*, n can be estimated from (14.13). We see from Table 14.6 that $n << 1$ for Mercury and Mars. A very large value of $n = 84$ is obtained for Venus. The large greenhouse effect for Venus is attributable primarily to CO_2, which constitutes in excess of 95% of the atmospheric mass. Earth has a moderate value of n =0.65.

At the very high temperature and pressure that characterize the Venusian surface, chemical reactions between the atmosphere and the surface become very important. It appears that the abundances of all of the reactive gases on Venus, including oxygen, carbon dioxide, and the sulfur compounds, are regulated primarily by buffer reactions with crustal minerals. A crustal source of CO_2 is of particular interest in the context of the greenhouse effect. The surface reactions appear to be dominated by the thermal reactions of sulfphur- and carbon-containing species. The Venusian surface contains a considerable amount of calcium carbonate ($CaCO_3$), which provides a source of atmospheric CO_2 at high temperatures and pressures. As an increasing amount of CO_2 becomes incorporated into the atmosphere, the greenhouse effect is enhanced, which increases the planetary temperature further and increases the breakdown of the lithospheric carbonates. This positive feedback between increasing surface temperatures and increasing concentration of greenhouse gases is known as the *runaway greenhouse effect*.

14.3.2 Planetary Time Scales

If radiative transfer is the only process occurring in a planetary atmosphere, the surface temperatures of the planets would be determined solely by the net radiation at the top of the atmosphere. Thus, the equatorial regions would be warmest and the poles would be extremely cold. Whereas this is the case on Mars and to a lesser extent on Earth, other planets such as Venus and all of the

Jovian planets have little or no equator-to-pole temperature gradients at the surface or deep in their interiors.

Simple arguments allow us to understand why there are differences in horizontal temperature gradients among the planets. There are two fundamental time scales that determine how a planetary atmosphere transfers heat. The *radiative timescale*, τ_{rad}, is the time it would take for an atmosphere above some pressure level p_0 to reduce its temperature by $1/e$ of its initial value via radiative cooling if solar radiation were turned off. The *dynamic timescale*, τ_{dyn}, is the time required to move a parcel over a characteristic distance in the atmosphere and, in so doing, transport heat from one location to another.

We can obtain a simple measure of a planetary radiative time scale as follows. Consider an atmosphere with a surface pressure of p_0. If the solar heating is turned off, the atmosphere will cool by thermal radiation to space. From (3.34), we can write the following expression for radiative heating

$$\frac{\partial \mathrm{T}}{\partial t} = \frac{g}{c_p}\frac{\partial F}{\partial p}$$

A simple estimate of the column radiative cooling for a black atmosphere can be obtained by using the slab model described in section 14.3.1. Assume that the entire column is initially in radiative equilibrium. At the instant the Sun's heating is turned off, each of the n slabs remains in radiative equilibrum except for the top slab, $n = 1$. Once the top slab has cooled, the remainder of the slabs cool by coming into radiative equilibrum with the top slab. Hence the flux divergence at the top of the atmosphere controls the cooling of the column, and we can write

$$\frac{\partial \mathrm{T}}{\partial t} = \frac{g}{c_p}\frac{\left(0-\sigma T_e^{*4}\right)}{p_0} = -\frac{g}{c_p}\frac{\sigma T_e^{*4}}{p_0} \qquad (14.14a)$$

where T_e* is the equivalent black body temperature of the planet. If the atmosphere is not black and $n < 1$, such as on Earth and Mars, then the atmosphere can cool more rapidly, since the entire atmosphere is cooling to space. In such a case, we increase the cooling rate relative to (14.11a) by scaling the surface pressure by n

$$\frac{\partial \mathrm{T}}{\partial t} = -\frac{g}{c_p}\frac{\sigma T_e^{*4}}{n\, p_0} \qquad (14.14b)$$

The product np_0 can be considered as the width of a weighting function for the atmospheric transmission, which describes the pressure-depth of the atmosphere over which most of the emission to space occurs. To determine the radiative time scale, we integrate (14.11) from T_e^* to $T_e^{*\prime} = T_e^*/e$, where $T_e^{*\prime}$ is an e-folding temperature for the cooling. Thus we obtain

$$\tau_{rad} = \frac{p_0 c_p}{g \sigma} \int_{T_e^{*\prime}}^{T_e^*} \frac{1}{T^4} dT = \frac{p_0 c_p}{4 g \sigma T_e^{*3} \left(1 - \frac{1}{e^3}\right)} \approx \frac{p_0 c_p}{4 g \sigma T_e^{*3}} \qquad (14.15a)$$

or if n < 1, we have

$$\tau_{rad} = \frac{n\, p_0 c_p}{4 g \sigma T_e^{*3}} \qquad (14.15b)$$

From (14.15a,b) it is possible to identify the planetary properties that determine the radiative time scale. The value of τ_{rad} decreases with increasing mass of the planet through g, but increases with increasing mass of the atmoshere through p_0. For a given mass of atmosphere (p_0), τ_{rad} decreases as a function of the temperature of the radiating temperature, reflecting the fourth-power black body radiative flux dependency on temperature. Finally, τ_{rad} depends on the composition of the atmosphere through the heat capacity c_p. All other things being equal, the radiative timescale for a hydrogen-helium atmosphere ($c_p \approx$ 12,000 J kg^{-1} K^{-1}) will be about an order of magnitude greater than for an atmosphere dominated by CO_2 or N_2 ($c_p \approx$ 1,000 J kg^{-1} K^{-1}).

If r_p is the planetary radius and u is a typical horizontal wind speed, the the dynamic time scale may be defined as

$$\tau_{dyn} = \frac{r_p}{u} \qquad (14.16)$$

The ratio of the two timescales, ε, is given by

$$\varepsilon = \frac{4 r_p g \sigma T_e^{*3}}{u p_0 c_p}, \quad n \geq 1 \qquad (14.17a)$$

or

$$\varepsilon = \frac{4 r_p g \sigma T_e^{*3}}{u n p_0 c_p}, \quad n < 1 \qquad (14.17b)$$

which provides a measure of the relative importance of radiative and dynamical effects.

Values of the time scales for each of the planets with a significant atmosphere are given in Table 14.7. There are several different regimes for ε.

(i) If ε >> 1, the dynamic time scale is much greater than the radiative time scale, and hence radiative processes dominate. This regime is characteristic of Mars. This regime occurs if u and/or p_0 are very small, reflecting slow planetary motions and small heat capacity, respectively. Thus, the fluxes of heat by horizontal atmospheric motions would be small compared to the local cooling rate by radiative effects. With ε >> 1, a parcel is in approximately in local radiative equibrium with outer space so that the temperature of the parcel during its motion would be determined by the local radiative fluxes. As a parcel is advected, it cools rapidly to space so that its initial temperature signature is rapidly lost. Consequently, on a planet with ε >> 1, the equator-to-pole temperature gradient is very large, in response the latitudinal variations of solar radiation. Also, as the radiative cooling of the atmospheric parcel is so efficient, a very large diurnal variation of temperature is expected.

(ii) If ε << 1, the dynamical transports of heat dominate over radiative cooling. This regime, characteristic of Venus, Jupiter, and Saturn, occurs because u is very large, indicating rapid transports of heat by horizontal motions, and p_0 is large, indicating very large atmospheric heat storage. Consider again the motion of a parcel from one point on a planet to another. Since the atmospheric pressure is so high, the time to cool the parcel radiatively is very long. With a slow radiative cooling rate, the temperature of the parcel varies adiabatically, retaining its initial thermal signature. Thus, for ε << 1 a very small or negligible equator-to-pole temperature difference is expected. For similar reasons, the diurnal variation of temperature on such a planet would be negligibly small.

(iii) For ε ~ 1, there is parity between the two time scales. Such a situation occurs if the mass of the atmosphere is moderate and the velocities on the planet are realitively weak, which is the case for Earth. Under these conditions, a parcel moving from one location to another will radiatively cool but not at a rate fast enough to destroy the characteristics of its initial thermal properties. On such a planet, one would expect a pole-to-equator temperature difference but one where the temperature of a parcel at a particular latitude is not in radiative equilibrium with the net radiative fluxes at the top of the atmosphere. Furthermore, a moderate diurnal temperature is expected.

Table 14.7: Estimates of the planetary radiation and dynamic timescales and their ratio ε for planets in the solar system. Estimates are for the entire atmospheres of the planets except for Jupiter and Saturn where only the upper 1 atmosphere was considered. A value of C_p of 12,000 J K^{-1} kg^{-1} is used for Jupiter and Saturn.

Planet	T_e (K)	p_0 (atm)	n	τ_{rad}	u (ms^{-1})	τ_{dyn}	ε
Venus	214	90	>1	36 y	2	35 d	0.003
Earth	254	1	0.65	18 d	10	7 d	0.39
Mars	217	0.01	0.12	0.05 d	5	10 d	200
Jupiter	110	1	>1	60 y	50	16 d	0.0007
Saturn	80	1	>1	299 y	50	14 d	0.0001

Venus has a planetary albedo of about 71% (Table 14.6) which is more than twice that of Earth. It is estimated that 24% of the incoming energy is absorbed in the cloud layer. The remaining energy penetrates the cloud layers but perhaps only 2% of the incident radiation at the top of the atmosphere reaches the surface. Observations indicate that the surface temperature of Venus is essentially the same at all latitudes and between the day and night siudes of the planet, at about 740 K. In the lower atmosphere of Venus $\varepsilon << 1$, due principally to the very large surface pressure. The vast heat capacity of the lower troposphere results in a radiative time scale that is orders of magnitude larger than the dynamic timescale (Table 14.7). Thus the motion of the atmopshere tends to homogenize the temperatures latitudinally and between the day and night sides of the planet.

For Earth, $\varepsilon \approx 0.4$. The mass of the Earth's atmosphere is sufficently large that the radiative cooling time scale is nearly equivalent to the dynamic time scale. Whereas the strong pole-to-equator temperature gradient suggests that radiative effects are very important, dynamic transports of heat tend to predominate. For example, the latitudinal surface temperature gradient does not follow the net radiation at the top of the atmosphere exactly, suggesting that transports of heat continually modify the surface temperatures. Occaisionally, dynamical transports of heat far exceed radiative effects. Each winter there are the familiar outbreaks of Arctic air masses that propagate to low latitudes before being modified by radiative heating and surface effects. Within the Earth's oceans, $\varepsilon << 1$ because of the high density and heat capacity of sea water. Since $\varepsilon << 1$, heat cannot be carried away in the ocean more quickly by radiative processes at the top of the column than can be transported by the deep ocean currents. As a result, the deep ocean is relatively homogeneous in temperature.

For Mars $\varepsilon >> 1$ principally because of the very low mass of the atmosphere, indicating that radiative processes dominate the dynamical transports of heat. As a result very large pole-to-equator temperature gradients are observed on the planet, as well as a strong diurnal cycle.

Estimates of the interior temperatures of Jupiter and Saturn indicate that there is little, if any, temperature gradient with latitude as one goes deeper into the interior of the atmosphere. This is a characteristic of all of the outer planets and is a further indication of extremely long radiative time scales compared to dynamic time scales. Thus, from (14.16a), $\varepsilon \ll 1$ for both Jupiter and Saturn.

Like Jupiter and Saturn, dynamic processes on Uranus and Neptune control temperature distributions deep within the atmospheres. The rotational axis of Uranus is in the solar plane so that one pole of the planet points towards the Sun for over 50 years at a time. Yet, the temperature of the planet appears to be relatively homogeneous within the interior, indicating once again that in dense planetary atmospheres that the radiative time scales are extremely long and the temperature distribution is dominated by dynamic processes.