

Chapter 2

The Global Energy Balance

NON_PRINT ITEMS

Abstract

In this chapter the key role of the energy balance in determining the climate of Earth is introduced. The emission temperature of a planet and its dependence on planetary albedo and total solar irradiance are described. Simple models of the greenhouse effect are applied. Formulas for the distribution of daily-averaged insolation as a function of latitude and season are derived. Geographic and seasonal variations of the Earth's top-of-atmosphere radiation balance are shown and discussed. The poleward transport of energy by the ocean and atmosphere are derived from the top-of-atmosphere energy constraint.

Key Words

Climate; energy budget of Earth; emission temperature of a planet; greenhouse effect; insolation; solar zenith angle; poleward energy transport; .

Chapter Starts here

2.1 Warmth and Energy

Temperature, a key climate variable, is a measure of the energy contained in the movement of molecules. To understand how the temperature is maintained, one must therefore consider the *energy balance* that is formally stated in the First Law of Thermodynamics. The basic global energy balance of Earth is between energy coming from the sun and energy returned to space by Earth's radiative emission. The generation of energy in the interior of Earth has a negligible influence on its energy budget. The absorption of solar radiation takes place mostly at the surface

of Earth, whereas most of the emission to space originates in its atmosphere. Because the atmosphere is mostly transparent to solar radiation and mostly opaque to terrestrial emission of radiation, the surface of Earth is much warmer than it would be in the absence of its atmosphere. When averaged over a year, more solar energy is absorbed near the equator than near the poles. The atmosphere and the ocean transport energy poleward to reduce the effect of this heating gradient on surface temperature. Much of the character of Earth's evolution and climate has been determined by its position within the solar system.

2.2 The Solar System

The source for the energy to sustain life on Earth comes from the sun. Earth orbits the sun once a year while keeping a relatively constant distance from it, so that our sun provides a stable, comfortable source of heat and light. Our sun is one of about 10^{11} stars in our galaxy, the Milky Way (Table 2-1). It is a single star, whereas about two-thirds of the stars we can see are in multiple star systems.

**** insert Table 2-1 near here *****

The luminosity is the total rate at which energy is released by the sun. We know of stars that are 10^{-4} times less luminous and 10^5 times brighter than the sun. Their emission temperatures range from 2000 to 30,000 K, whereas the temperature of the photosphere of the sun is about 6000 K. The photosphere is the region of the sun from which most of its energy emission is released to space. Stellar radii range from 0.1 to 200 solar radii. Energy is produced in the core of the sun by nuclear fusion, whereby lighter elements are made into heavier ones, releasing energy in the process. For a smallish star such as the sun, the projected lifetime on the main sequence is about 11 billion years, of which about half has passed. The sun is thus a solitary, middle-aged, medium-bright star. Theories of stellar evolution predict that the luminosity of the sun has

increased by about 30% during the lifetime of Earth, about 4.5 billion years. The cause of this increased energy output is the gradual increase in the density of the sun, as lighter elements convert to heavier ones.

The solar system includes nine planets. They may be divided into the *terrestrial*, or inner, planets and the *Jovian*, or outer, planets (Table 2-2). The terrestrial planets include Mercury, Venus, Mars, and Earth. The Jovian planets include Jupiter, Saturn, Uranus, and Neptune. Pluto doesn't fit neatly into either category (Table 2-3).

*** Insert Table 2-2 near here ***

*** Insert Table 2-3 near here ***

2.2.1 Planetary Motion

The planets orbit about the sun in ellipses, which have three characteristics: the mean planet-sun distance, the eccentricity, and the orientation of the orbital plane. The mean distance from the sun controls the amount of solar *irradiance* (energy delivered per unit time per unit area) arriving at the planet. The mean distance from the sun also controls the length of the planetary year, the time it takes the planet to complete one orbit. The planetary year increases with increasing distance from the sun.

The *eccentricity* of an orbit is a measure of how much the orbit deviates from being perfectly circular. It controls the amount of variation of the solar irradiance at the planet as it moves through its orbit during the planetary year. If the eccentricity is not zero, so that the orbit is not circular, then the distance of the planet from the sun varies through the year. The orientation of the orbital plane doesn't have too much direct bearing on the climate. Most planets except Pluto

are in more or less the same orbital plane, which is perpendicular to the axis of the Sun's rotation, because the solar system formed from a spinning cloud of gas and rock.

In addition to the orbit parameters, the parameters of the planets' rotation and their relationship to the orbit are very important. The rotation rate controls the temporal behavior of the insolation at a point (diurnal cycle) and is also an important control on the response of the atmosphere and ocean to solar heating and thereby on the patterns of winds and currents that develop.

The *obliquity* or tilt is the angle between the axis of rotation and the normal to the plane of the orbit. It influences the seasonal variation of insolation, particularly in high latitudes. It also strongly affects the annual mean insolation that reaches the polar regions. Currently the obliquity of Earth's axis of rotation is about 23.45° .

The longitude of perihelion measures the phase of the seasons relative to the planet's position in the orbit. For example, at present Earth passes closest to the sun (*perihelion*) during Southern Hemisphere summer, on about January 5. As a result the Southern Hemisphere receives more top-of-atmosphere insolation than the Northern Hemisphere.

The effect of these orbital parameters on climate will be discussed in more detail in Chapter 12, where the orbital parameter theory of climate change is described. For the time being we consider only the distance from the sun and the obliquity or declination angle.

2.3 Energy Balance of Earth

2.3.1 First Law of Thermodynamics

The First Law of Thermodynamics states that energy is conserved. The first law for a closed system may be stated as: "The heat added to a system is equal to the change in internal energy minus the work extracted." This law may be expressed symbolically as

$$dQ = dU - dW \quad (2.1)$$

Here dQ is the amount of heat added, dU is the change in the internal energy of the system, and dW is the work extracted from the system. Heat can be transported to and from a system in three ways.

1. *Radiation*: No mass is exchanged, and no medium is required. Pure radiant energy moves at the speed of light.
2. *Conduction*: No mass is exchanged, but a medium transfers heat by collisions between atoms or molecules.
3. *Convection*: Mass is exchanged. A net movement of mass may occur, but more commonly parcels with different energy amounts change places, so that energy is exchanged without a net movement of mass (Fig. 2-1).

*** insert Figure_2-1 near here ***

The transmission of energy from the sun to Earth is almost entirely radiative. Some mass flux is associated with the particles of the solar wind, but the amount of energy is small compared to the energy arriving as photons. Moreover, the work done by Earth on its environment is also negligible. To calculate the approximate energy balance of Earth, we need only consider radiative energy exchanges. The amount of matter in space that could affect the flow of energy between the sun and Earth is small, and we may thus consider the space between the photosphere of the sun and the top of Earth's atmosphere to be a vacuum. In a vacuum, only radiation can transport energy.

2.3.2 Energy Flux, Flux Density, and Solar Constant

The sun puts out a nearly constant flux of energy that we call the solar *luminosity*, $L_0 = 3.9 \times 10^{26}$ Watts. We can calculate the average irradiance at the photosphere by dividing this energy flux by the area of the photosphere.

$$\begin{aligned} \text{Irradiance}_{\text{photo}} &= \frac{\text{flux}}{\text{area}_{\text{photo}}} = \frac{L_0}{4\pi r_{\text{photo}}^2} \\ &= \frac{3.9 \times 10^{26} \text{ W}}{4\pi [6.96 \times 10^8 \text{ m}]^2} = 6.4 \times 10^7 \text{ W m}^{-2} \end{aligned} \quad (2.2)$$

Since space is effectively a vacuum and energy is conserved, the amount of energy passing outward through any sphere with the sun at its center should be equal to the luminosity, or total energy flux from the sun. If we assume that the irradiance is uniform over the sphere, and write the irradiance at any distance d from the sun as S_d , then conservation of energy requires

$$\text{Flux} = L_0 = S_d 4\pi d^2 \quad (2.3)$$

From this we deduce that the total solar irradiance S_d is inversely proportional to the square of the distance to the sun. We define the *total solar irradiance* (TSI) as the irradiance of the solar emission at the mean distance of the Earth from the Sun.

$$\text{TSI} = \text{irradiance at distance } d = S_d = \frac{L_0}{4\pi d^2} \quad (2.4)$$

The measurement of the total solar irradiance has a long and interesting history, including trying to estimate it from the reflection from the Moon. The most reliable estimates come from special instruments placed on Earth-orbiting satellites, so that a direct measurement of the irradiance can be made without an intervening atmosphere. These instruments have achieved great precision, so that the variability of the TSI on periods of hours to decades has been measured with great reproducibility by different instruments, but the mean absolute value has

varied from instrument to instrument. The best and most recent estimate of the TSI is $1360.8 \pm 0.5 \text{ Wm}^{-2}$.¹ For the purposes of doing exercises in this book we will assume that the TSI is 1360 Wm^{-2} at the mean distance of Earth from the sun ($1.5 \times 10^{11} \text{ m}$).

2.3.3 Cavity Radiation

The radiation field within a closed cavity in thermodynamic equilibrium has a value that is uniquely related to the temperature of the cavity walls, regardless of the material of which the cavity is made. This cavity radiant intensity, which is uniquely related to the wall temperature, is also called the *blackbody radiation*, since it corresponds to the emission from a surface with unit emissivity and absorptivity. Perfect blackbodies may not be easily found, but the radiation inside a cavity in equilibrium will always equal the blackbody radiation. The dependence of the blackbody emission on temperature follows the Stefan–Boltzmann Law.

$$E_{BB} = \sigma T^4; \quad \sigma = 5.67 \times 10^{-8} \text{ Wm}^{-2} \text{ K}^{-4} \quad (2.5)$$

2.3.4 Example: Emission Temperature of the Sun

We calculated earlier that the solar irradiance at the photosphere is about $6.4 \times 10^7 \text{ W m}^{-2}$.

We can equate this to the Stefan–Boltzmann formula (2.5) and derive an effective emission temperature for the photosphere.

$$\begin{aligned} \sigma T_{\text{photo}}^4 &= 6.4 \times 10^7 \text{ W m}^{-2} \\ T_{\text{photo}} &= \sqrt[4]{\frac{6.4 \times 10^7 \text{ W m}^{-2}}{\sigma}} = 5796 \text{ K} \sim 6000 \text{ K} \end{aligned}$$

¹ Kopp and Lean, 2011

2.3.5 Emissivity

In equilibrium, the radiant intensity inside a cavity at temperature T is $E_{BB} = \sigma T^4$. We may define the emissivity, ε , as the ratio of the actual emission of a body or volume of gas to the blackbody emission at the same temperature.

$$E_R = \varepsilon \sigma T^4 \rightarrow \varepsilon = \frac{E_R}{\sigma T^4} \quad (2.6)$$

2.4 Emission Temperature of a Planet

The emission temperature of a planet is the blackbody temperature at which it needs to emit in order to achieve energy balance. The basic idea is to equate the solar energy absorbed by a planet with the energy emitted by a blackbody. This defines the emission temperature of the planet.

$$\text{Solar radiation absorbed} = \text{planetary radiation emitted}$$

To calculate the solar radiation absorbed we begin with the total solar irradiance, which measures the energy flux of solar radiation in Watts per square meter (Wm^{-2}) arriving at the mean distance of the planet from the sun. The energy flux is defined relative to a flat surface perpendicular to the direction of radiation. Solar radiation is essentially a parallel and uniform beam for a planetary body in the solar system, because the planets all have diameters that are small compared to their distance from the sun. The amount of energy incident on a planet is equal to the TSI times the area that the planet sweeps out of the beam of parallel energy flux. We call this the *shadow area* (Fig. 2-2). Since the atmosphere of Earth is very thin, we can ignore its effect on the shadow area and use the radius of the solid planet, r_p , to calculate the shadow area.

*** Insert Figure_2-1 near here ***

We must also take account of the fact that not all the solar energy incident on a planet is absorbed. Some fraction is reflected back to space without being absorbed and so does not enter

into the planetary energy balance. We call this planetary reflectivity the planetary albedo¹ and give it the symbol α_p . Thus we have

$$\text{Absorbed solar radiation} = S_0(1 - \alpha_p)\pi r_p^2 \quad (2.7)$$

Since the *planetary albedo* for Earth is about 29%, only 71% of the TSI is absorbed by the climate system. This amount of energy must be returned to space by terrestrial emission. We assume that the terrestrial emission is like that of a blackbody. The area from which the emission occurs is the surface area of a sphere, rather than the area of a circle. The terrestrial emission flux is thus written

$$\text{Emitted terrestrial radiation} = \sigma T_e^4 4\pi r_p^2 \quad (2.8)$$

If we equate the absorbed solar flux with the emitted terrestrial flux, we obtain the planetary energy balance, which will define the emission temperature.

$$\frac{S_0}{4}(1 - \alpha_p) = \sigma T_e^4 \quad (2.9)$$

or

$$T_e = \sqrt[4]{\frac{(S_0/4)(1 - \alpha_p)}{\sigma}} \quad (2.10)$$

The factor of 4 dividing the total solar irradiance in (2.9) is the ratio of the global surface area of a sphere to its shadow area, which is the area of a circle with the same radius. The emission temperature may not be the actual surface or atmospheric temperature of the planet; it is merely the blackbody emission temperature a planet requires to balance the solar energy it

¹ Albedo comes from a Latin word for “whiteness.”

absorbs. For a planet with a strong greenhouse effect like Earth, most of the emission comes from the atmosphere, not the surface.

2.4.1 Example: Emission Temperature of Earth

Earth has an albedo of about 0.29. The emission temperature of Earth from (2.10) is therefore

$$T_e = \sqrt[4]{\frac{(1360 \text{ W m}^{-2}/4)(1-0.3)}{5.67 \times 10^{-8} \text{ W m}^{-2}\text{K}^{-4}}} \cong 255\text{K} \cong -19^\circ\text{C} \cong -2^\circ\text{F}$$

The emission temperature of 255 K is much less than the observed global mean surface temperature of 288 K $\cong +15^\circ\text{C}$. To understand the difference we need to consider the greenhouse effect.

2.5 Greenhouse Effect

One may illustrate the greenhouse effect with a very simple elaboration of the energy balance model used to define the emission temperature. An atmosphere that is assumed to be a blackbody for terrestrial radiation, but is transparent to solar radiation, is incorporated into the global energy balance (Fig. 2-3). Since solar radiation is mostly visible and near infrared, and Earth emits primarily thermal infrared radiation, the atmosphere may affect solar and terrestrial radiation very differently. The energy balance at the top of the atmosphere in this model is the same as in the basic energy balance model that defined the emission temperature (2.9). Since the atmospheric layer absorbs all of the energy emitted by the surface below it and emits like a blackbody, the only radiation emitted to space is from the atmosphere in this model. The energy balance at the top of the atmosphere is thus

$$\frac{S_0}{4}(1 - \alpha_p) = \sigma T_A^4 = \sigma T_e^4 \quad (2.11)$$

Therefore we see that the temperature of the atmosphere in equilibrium must be the emission temperature in order to achieve energy balance. The surface temperature is much warmer,

however, as we can see by deriving the energy balance for the atmosphere and the surface. The atmospheric energy balance gives,

$$\sigma T_s^4 = 2\sigma T_A^4 \Rightarrow \sigma T_s^4 = 2\sigma T_e^4 \quad (2.12)$$

and the surface energy balance is consistent:

$$\frac{S_0}{4}(1 - \alpha_p) + \sigma T_A^4 = \sigma T_s^4 \Rightarrow \sigma T_s^4 = 2\sigma T_e^4 \quad (2.13)$$

From (2.13) we obtain $T_s = \sqrt[4]{2} T_e = 303K$, which is warmer than the observed global mean surface temperature. This is because we have not taken into account the vertical transport of energy by atmospheric motions.

*** Insert Figure_2-3 here ***

We can see from the diagram in Fig. 2-3 and the surface energy balance (2.13), that the surface temperature is increased because the atmosphere does not inhibit the flow of solar energy to the surface, but augments the solar heating of the surface with its own downward emission of longwave radiation. In this model the surface receives the same amount of energy as downward emission of terrestrial radiation from the atmosphere as it receives from the sun. The atmospheric greenhouse effect warms the surface because the atmosphere is relatively transparent to solar radiation and yet absorbs and emits terrestrial radiation very effectively.

2.6 Global Radiative Flux Energy Balance

The vertical flux of energy in the atmosphere is one of the most important climate processes. The radiative and non-radiative fluxes between the surface, the atmosphere, and space are key determinants of climate. The ease with which solar radiation penetrates the atmosphere and the difficulty with which terrestrial radiation is transmitted through the atmosphere determine the

strength of the greenhouse effect. The decrease of temperature with altitude (lapse rate) is also a key part of the greenhouse effect.

The fundamental energy exchanges in the climate system at the top of the atmosphere, within the atmosphere and at the surface are shown in Fig. 2-4. At the top of the atmosphere absorbed solar radiation is about 240 Wm^{-2} and emitted terrestrial radiation is about 239 Wm^{-2} . The difference of about 0.6 Wm^{-2} is being stored in the ocean, which is heating up at the present time, very likely as a result of human production of greenhouse gases. More than 90% of the terrestrial emission to space originates in the atmosphere and only about 20 Wm^{-2} comes from the surface through the atmospheric infrared window. Of the 240 Wm^{-2} of solar radiation absorbed in the climate system, only about a third is absorbed in the atmosphere and two thirds is absorbed at the surface. Although the surface receives 160 Wm^{-2} of solar radiation, this is less than half of the 345 Wm^{-2} that the surface receives as downward thermal emission from the atmosphere. The strong downward longwave radiation from the atmosphere is also important in limiting the diurnal variation of surface temperature, which might otherwise cool rapidly at night. The surface emits 396 Wm^{-2} of radiation upward, so that net sum of upward and downward terrestrial radiation cools the surface at a net rate of 51 Wm^{-2} , which is smaller than the sum of evaporation (88 Wm^{-2}) and convection of warmth away from the surface (20 Wm^{-2}).

*** insert Figure_2-4 near here ***

One way to measure the strength of the greenhouse effect is to compare the emission from the surface with the emission from the top of the atmosphere. These differ by $157 \text{ Wm}^{-2} = 396 \text{ Wm}^{-2} - 239 \text{ Wm}^{-2}$, which is how much the atmosphere decreases the emission of energy to space for the current climate. This depends not only on the opacity of the atmosphere for longwave radiation, but also on the transparency of the atmosphere to solar radiation and the fact

that the temperature of the atmosphere decreases with altitude. The lapse rate of temperature is possible because the atmosphere is a compressible gas.

Using the numbers in Fig. 2-4, one can construct the energy budgets at the top of atmosphere, the atmosphere and the surface.

Top of atmosphere:

Incoming solar – reflected solar - emitted terrestrial radiation = storage

$$340 \text{ Wm}^{-2} - 100 \text{ Wm}^{-2} - 239 \text{ Wm}^{-2} = 0.6 \text{ Wm}^{-2}$$

Atmosphere:

Absorbed solar + Thermals + latent heating + surface longwave – outgoing longwave = 0

$$80 \text{ Wm}^{-2} + 20 \text{ Wm}^{-2} + 88 \text{ Wm}^{-2} + 51 \text{ Wm}^{-2} - 239 \text{ Wm}^{-2} = 0$$

The atmospheric heat budget can be rearranged in the following way,

absorbed solar + surface longwave – outgoing longwave

= Net radiative cooling = thermal + condensation heating

$$80 \text{ Wm}^{-2} + 51 \text{ Wm}^{-2} - 239 \text{ Wm}^{-2} = 108 \text{ Wm}^{-2} = 20 \text{ Wm}^{-2} + 88 \text{ Wm}^{-2}$$

So that 81% of the radiative cooling of the atmosphere is balanced by condensational heating.

Surface:

Absorbed solar - Thermals - latent heating - surface longwave = Storage

$$160 \text{ Wm}^{-2} - 20 \text{ Wm}^{-2} - 88 \text{ Wm}^{-2} - 51 \text{ Wm}^{-2} = 0.6 \text{ Wm}^{-2}$$

The very strong downward emission of terrestrial radiation from the atmosphere is essential for maintaining the relatively small diurnal variations in surface temperature over land.

If the downward longwave were not larger than the solar heating of the surface, then the land

surface temperature would cool very rapidly at night and would warm exceedingly rapidly during the day, yielding a large diurnal variation of surface temperature.

2.7 Distribution of Insolation

Insolation is the amount of downward solar radiation energy incident on a plane surface. Seasonal and latitudinal variations in temperature are driven primarily by variations of insolation and average solar zenith angle. The amount of solar radiation incident on the top of the atmosphere depends on the latitude, season, and time of day. The amount of solar energy that is reflected to space without absorption depends on the solar zenith angle and the properties of the local surface and atmosphere. The climate depends on the insolation and zenith angle averaged over a 24-hour period, over a season, and over a year. In this section the geometric factors that determine insolation and solar zenith angle will be described. The 1360 Wm^{-2} mean solar flux per unit area at the mean position of Earth is measured for a surface that is perpendicular to the solar beam. Because Earth is approximately spherical, virtually all of the planet's sunlit surface is inclined at an oblique angle to the solar beam. Unless the sun is directly overhead, the solar irradiance is spread over a surface area that is larger than the perpendicular area, so that the irradiance per unit surface area is smaller than the TSI. We define the *solar zenith angle*, θ_s , as the angle between the local normal to Earth's surface and a line between a point on Earth's surface and the sun. Figure 2-5 indicates that the ratio of the shadow area to the surface area is equal to the cosine of the solar zenith angle. We may write the solar flux per unit surface area as

$$Q = S_0 \left(\frac{\bar{d}}{d} \right)^2 \cos \theta_s \quad (2.14)$$

where \bar{d} is the mean distance for which the TSI S_0 is measured, and d is the actual distance from the sun.

** Insert Figure_2-5 near here **

The solar zenith angle depends on the latitude, season, and time of day. The season can be expressed in terms of the *declination angle* of the sun, which is the latitude of the point on the surface of Earth directly under the sun at noon. The declination angle (δ) currently varies between $+23.45^\circ$ at northern summer solstice (June 21) to -23.45° at northern winter solstice (December 21). The *hour angle*, h , is defined as the longitude of the sub solar point relative to its position at noon. If these definitions are made, then the cosine of the solar zenith angle can be derived for any latitude (ϕ), season, and time of day from spherical trigonometry formulas (see Appendix A).

$$\cos \theta_s = \sin \phi \sin \delta + \cos \phi \cos \delta \cos h \quad (2.15)$$

If the cosine of the solar zenith angle is negative, then the sun is below the horizon and the surface is in darkness. Sunrise and sunset occur when the solar zenith angle is 90° , in which case (2.15) gives

$$\cos h_0 = -\tan \phi \tan \delta \quad (2.16)$$

where h_0 is the hour angle at sunrise and sunset. At noon the solar zenith angle depends only on the latitude and declination angle, $\theta_s = \phi - \delta$.

Near the poles special conditions prevail. When the latitude and the declination angle are of the same sign (summer), latitudes poleward of $90-\delta$ are constantly illuminated. At the pole the sun moves around the compass at a constant angle of δ above the horizon. In the winter hemisphere, where ϕ and δ are of the opposite sign, latitudes poleward of $90-|\delta|$ are in polar darkness. At the poles, six months of darkness alternate with six months of sunlight. At the equator day and night are both twelve hours long throughout the year. The average daily

insolation on a level surface at the top of the atmosphere may be obtained by substituting (2.15) into (2.14), integrating the result between sunrise and sunset, and then dividing by 24 hours. The result is

$$\bar{Q}^{day} = \frac{S_0}{\pi} \left(\frac{\bar{d}}{d} \right)^2 [h_0 \sin \phi \sin \delta + \cos \phi \cos \delta \sin h_0] \quad (2.17)$$

where the hour angle at sunrise and sunset, h_0 , must be given in radians. The daily average insolation is plotted in Fig. 2-6 as a function of latitude and season. Earth's orbit is not exactly circular, and currently Earth is somewhat closer to the sun during Southern Hemisphere summer than during Northern Hemisphere summer. As a result the maximum insolation in the Southern Hemisphere is about 6.9% higher than that in the Northern Hemisphere. Note that at the summer solstice the insolation in high latitudes is actually greater than that near the equator. This results from the very long days during summer and in spite of the relatively large solar zenith angles at high latitudes.

*** Insert Figure_2-6 near here ***

If the daily insolation is averaged over the entire year, the distribution given in Fig. 2-7 is obtained. The annual-average insolation at the top of the atmosphere at the poles is less than half its value at the equator, where it reaches a maximum. By comparing the annual mean insolation at the equator with the insolation at the solstices one can see that the insolation at the equator goes through a semiannual variation with maxima at the equinoxes and minima at the solstices. This is because the sun crosses the equator twice a year at the equinoxes. Because the local albedo of Earth depends on the solar zenith angle, the zenith angle enters in determining both the available energy per unit of surface area and the albedo. It is therefore of interest to consider the average solar zenith angle during the daylight hours as a function of latitude and season. In calculating a daily average zenith angle it is appropriate to weight the average with respect to the

insolation, rather than time. A properly weighted average zenith angle is thus computed with the following formula,

$$\overline{\cos\theta_s}^{day} = \frac{\int_{-h_0}^{h_0} Q \cos\theta_s dh}{\int_{-h_0}^{h_0} Q dh} \quad (2.18)$$

where Q is the instantaneous insolation given by (2.14).

*** Insert Figure_2-7 near here ***

*** Insert Figure_2-8 near here ***

*** Insert Figure_2-9 near here ***

The daily average solar zenith angle calculated according to (2.18) varies from a minimum of 38.3° at the subsolar latitude and increases to 90° at the edge of the polar darkness (Figs. 2.8 and 2.9). At the pole the minimum value of the daily average solar zenith angle is achieved at the summer solstice, when it equals $\phi - \delta = 66.55^\circ$. Because of the much higher average solar zenith angles in high latitudes, more solar radiation is reflected than would be from a similar scene in tropical latitudes.

2.8 The Energy Balance at the Top of the Atmosphere

The amount of energy absorbed and emitted by Earth varies geographically and seasonally, depending on atmospheric and surface conditions as well as the distribution of insolation. The energy balance at the top of the atmosphere is purely radiative and can be measured accurately from Earth-orbiting satellites. The net radiation at the top of the atmosphere (R_{TOA}) is equal to the absorbed solar radiation (Q_{abs}) minus the outgoing longwave radiation (OLR).

$$R_{TOA} = Q_{abs} - OLR \quad (2.19)$$

The absorbed solar radiation is the incident solar radiation at the top of the atmosphere S_{TOA} minus the fraction reflected, which is the albedo, α .

$$Q_{abs} = S_{TOA}(1 - \alpha) \quad (2.20)$$

The albedo is estimated by measuring the solar radiation reflected from a region of Earth and comparing that with the insolation. The global mean, or planetary, albedo of Earth is about 29%. The albedo shows interesting geographic structure (Fig. 2-10).² It is highest in the polar regions where cloud and snow cover are plentiful and where average solar zenith angles are large. Secondary maxima of albedo occur in tropical and subtropical regions where thick clouds are prevalent or over bright surfaces such as the Sahara Desert. The smallest albedos occur over those tropical ocean regions where clouds are sparsely distributed. The ocean surface has an intrinsically low albedo, so that when clouds and sea ice are absent the Earth's albedo over ocean areas is only 8–10%.

*** Insert Figure 2-10 near here ***

Outgoing longwave radiation (OLR) is greatest over warm deserts and over tropical ocean areas where clouds occur infrequently (Fig. 2-11). It is lowest in polar regions and in regions of persistent high cloudiness in the Tropics. The temperature of the emitting substance controls the OLR, so that the cold poles and the cold cloud tops produce the lowest values. A warm surface that is overlain by a relatively dry, cloudless atmosphere produces the highest OLR values.

The net radiation is negative near the poles and positive in the Tropics (Fig. 2-12). The highest positive values of about 120 W m^{-2} occur over the subtropical oceans in the summer hemisphere, where large insolation and relatively low albedos both contribute to large absorbed

² We use a Hammer equal-area projection at many points in this book. In displaying irradiance (Wm^{-2}), it is especially important that the fraction of the area of the projection occupied by a particular feature be proportional to the fraction of the area of Earth's surface that the feature occupies.

solar radiation. The greatest losses of energy occur in the polar darkness of the winter hemisphere, where the OLR is uncompensated by solar absorption. Dry desert areas such as the Sahara in northern Africa are interesting because, despite their subtropical latitude, they lose energy in the annual average. Their energy loss is related to the relatively high surface albedo of dry deserts, combined with the large OLR loss associated with a dry, cloudless atmosphere above a warm surface. Since these areas are over land, which cannot transport heat laterally, the radiative loss out the top of the atmosphere must be balanced by horizontal heat transport by the atmosphere into the desert region from elsewhere.

*** Insert Figure 2-11 near here ***

*** Insert Figure 2-12 near here ***

When averaged around latitude circles, the components of the energy balance at the top of the atmosphere clearly show the influence of the latitudinal gradient of insolation (Fig. 2-13). The poleward decrease in absorbed solar energy is even greater than that of insolation because the albedo increases with latitude. Thus a smaller fraction of the available insolation is absorbed at higher latitudes than at the equator. The albedo increases with latitude because solar zenith angle, cloud coverage, and snow cover all increase with latitude. The energy emitted to space by the atmosphere does not decrease with latitude as rapidly as the absorbed solar radiation, because the atmosphere and ocean transport heat poleward and support a net energy loss to space in polar regions. The absorbed solar radiation exceeds the OLR in the tropics, so that the net radiation is positive there. Poleward of about 40° the absorbed solar radiation falls below the OLR and the net radiation balance is negative, so that the climate system loses energy to space. The latitudinal gradient in annual-mean net radiation must be balanced by a poleward flux of energy in the atmosphere and ocean.

*** Insert Figure 2-13 near here ***

*** insert Fig. 2-14 near here ***

*** insert Fig. 2-15 near here ***

2.9 Poleward Energy Flux

In the lowest curve of Fig. 2-13 we see that the annual mean net radiation is positive equatorward of about 40° of latitude and negative poleward of that latitude. As illustrated in Fig. 2-14, the energy balance for the climate system involves only the exchange at the top of the atmosphere, the transport through the lateral boundaries of the region in question by the atmosphere and ocean, and the time rate of change of energy within the region. Energy exchange with the solid earth can be neglected. We can write the energy balance for the climate system as

$$\frac{\partial E_{ao}}{\partial t} = R_{TOA} - \Delta F_{ao} \quad (2.21)$$

where $\partial E_{ao}/\partial t$ is the time rate of change of the energy content of the climate system, R_{TOA} is the net incoming radiation at the top of the atmosphere, and ΔF_{ao} is the divergence of the horizontal flux in the atmosphere and ocean. If we average over a year, then the storage term becomes small, and we have an approximate balance between net flux at the top of the atmosphere and horizontal transport.

$$R_{TOA} = \Delta F_{ao} \quad (2.22)$$

We can then use the observation of net radiation in Fig. 2-13 to derive the required annual-mean energy transport in the north-south direction. We assume that the energy lost to space poleward of any latitude must be supplied by poleward transport across that latitude by the

atmosphere and the ocean. So if we integrate the net radiation over a polar cap area, we can calculate the total energy flux across each latitude belt in the following way:

$$F(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} R_{TOA} a^2 \cos\phi \, d\lambda \, d\phi \quad (2.23)$$

The total northward energy flux inferred from the radiation imbalance peaks in midlatitudes at about 5.8×10^{15} W, or 5.8 petawatts (PW) (Fig. 2-15). This flux includes the contributions from both the atmosphere and the ocean. The flux of energy in the atmosphere can be estimated from analyzed fields of wind, temperature, and humidity. If this flux is subtracted from the total flux, it gives an estimate of the oceanic energy flux, which is more difficult to estimate from direct measurements. At 20°N, the contributions from the atmospheric and oceanic poleward fluxes are each about 2.0 PW. The ocean flux peaks at about 20°N in the subtropics, whereas the atmospheric flux has a broad maximum peaking near 40°N and 40°S. The oceanic heat flux is somewhat weaker in the Southern Hemisphere, but also peaks in the subtropics. In Chapters 6 and 7 we investigate in more detail how the atmosphere and the oceans accomplish this poleward flux of energy. If the fluid envelope of Earth did not transport heat poleward, the Tropics would be warmer and the polar regions would be much colder. Transport of heat by the atmosphere and the oceans makes the climate of Earth much more equable than it would otherwise be.

If we integrate over the globe, the horizontal transport is zero, since energy cannot be removed from spherical Earth through horizontal transport. In the annual global average then, the net radiation is very close to zero. Any global mean net radiative imbalance will lead to a heating or cooling of Earth.

Exercises

1. Use the data in Table 2.3 to calculate the emission temperatures for all of the planets. The actual energy emission from Jupiter corresponds to an emission temperature of about 124K. How must you explain the difference between the number you obtain for Jupiter and 124K?
2. Calculate the emission temperature of Earth if the solar luminosity is 30% less, as it is hypothesized to have been early in Solar System history. Use today's albedo and Earth-sun distance.
3. Calculate the emission temperature of Earth, if the planetary albedo is changed to that of ocean areas without clouds, about 10%.
4. Using the model illustrated in Fig. 2.3, calculate the surface temperature if the insolation is absorbed in the atmosphere, rather than at the surface.
5. Using the data in Fig. 2.4, estimate the blackbody temperature of Earth's surface and the blackbody temperature of the atmosphere as viewed from Earth's surface. How well do these temperatures agree with the temperatures derived from the model shown in Fig. 2.3?
6. Calculate the solar zenith angle at 9AM local sun time at Seattle (47°N , 122°W) at summer solstice and at winter solstice. Do the same for your hometown.
7. A mountain range is oriented east-west and has a slope on its north and south faces of 15 degrees. It is located at 45°N . Calculate the insolation per unit surface area on its south and north faces at noon at the summer and winter solstices. Compare the difference between the insolation on the north and south faces in each season with the seasonal variation on each face. Ignore eccentricity and atmospheric absorption.

8. In his science fiction novel, *Ringworld*, Larry Niven imagines a manufactured world consisting of a giant circular ribbon of superstrong material that circles a star. The width of the ribbon normal to a radius vector originating at the star's center is much greater than the thickness of the ribbon in the radial direction. The width of the ribbon is much less than the distance of the ribbon from the star, however. If the diameter of the ribbon is the same as the diameter of Earth's orbit about the sun, the luminosity is the same as Earth's sun, and the albedo of the ribbon is 0.3, what is the emission temperature of the sunlit side of the ribbon? Consider both the case in which heat is conducted very efficiently from the sunlit to the dark side of the ribbon, and the case in which no heat is conducted from the sunlit to the dark side of the ribbon. Explain why your emission temperature is different from Earth's. How big would the ribbon need to be to reproduce the emission temperature of Earth?