CHAPTER 2 What makes it go?

2.1 The Earth's radiation budget: An "upper boundary condition" on the general circulation

Radiation is (almost) the only mechanism by which the Earth can exchange energy with the rest of the Universe [1]. The solar energy flux at the mean radius of the Earth's orbit is about 1370 W m^{-2}. One way to get an intuitive grasp of this number is to imagine fourteen 100-Watt light bulbs per square meter. Another is to consider that 1370 W m^{-2} is the same as 1.37 GW km^{-2}, i.e., equivalent to the energy output of a large power plant for each square kilometer of area normal to the solar beam.

The globally averaged top-of-the-atmosphere radiation budget is summarized in Table 2.1. The Earth's albedo is near 0.30, independent of season; this number has been

Table 2.1: Summary of the annually averaged top-of-the-atmosphere radiation budget.

Incident solar radiation	340 W m^{-2}
Absorbed solar radiation	240 W m^{-2}
Planetary albedo	0.30
Outgoing longwave radiation	240 W m^{-2}
Brightness temperature of the Earth	255 K

known to better than 10% accuracy only since the 1970s. The energy absorbed by the Earth is

[1] Actually, there are two other mechanisms, namely gravitational interactions between the Earth and other objects (i.e., tidal effects), and the capture of extraterrestrial material by the Earth.

$$
\begin{aligned}
S_{abs} &= S\left(\frac{\pi a^2}{4\pi a^2}\right)(1-\alpha) \\
&= \frac{1}{4}S(1-\alpha) \cong 240\ \mathrm{W\ m^{-2}}(\text{annual mean})\ .
\end{aligned}
\tag{2.1}
$$

Here S_{abs} is the average absorbed solar energy per unit area. S is the "solar constant," a is the radius of the Earth, and α is the planetary albedo. Note that this rate of energy absorption is referred to the total surface area of the Earth, i.e. $4\pi a^2$, rather than to the cross-sectional or "projected" area presented to the solar beam, which is four times smaller.

The most important upper boundary condition on the general circulation of the atmosphere is the incident solar radiation. It is basically determined by the Earth's orbital parameters, (see Fig. 2.1) including the obliquity, eccentricity, and the dates of the

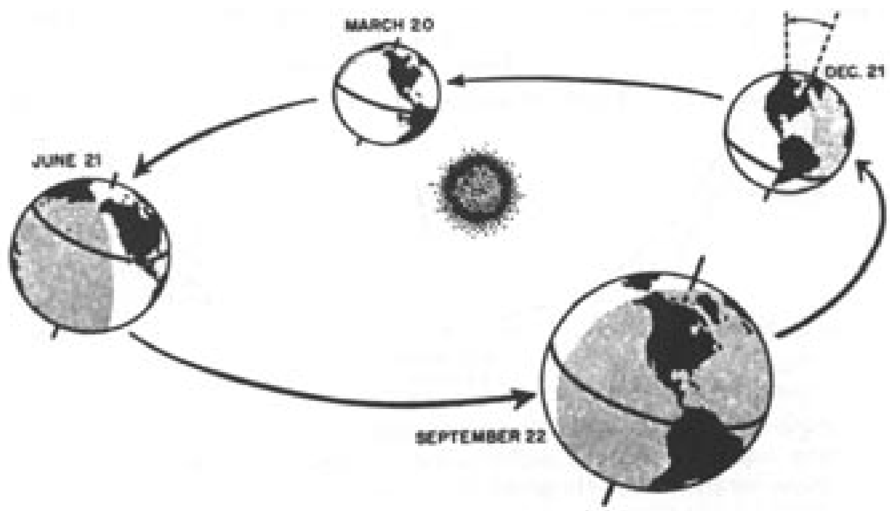

Figure 2.1: March of the seasons. As the tilted Earth revolves around the sun, changes in the distribution of sunlight cause the succession of seasons. (From Imbrie and Imbrie, 1979)

equinoxes, as well as the Earth's rotation rate. These all vary over geologic time (e.g. Crowley and North, 1991.

As a matter of common experience, the insolation varies both diurnally and seasonally. At a given moment, the insolation also varies strongly with longitude. Because a year is much longer than a day, the daily-mean insolation is (almost) independent of longitude, but it varies strongly with latitude in a way that depends on the season, as summarized in Fig. 2.2. As we move from the solar Equator to the summer pole, the insolation initially decreases, because at a given local time (e.g., local noon) the sun appears to be lower in the sky. In addition, however, the length of day increases at higher latitudes, and this obviously tends to make the daily-mean insolation increase.

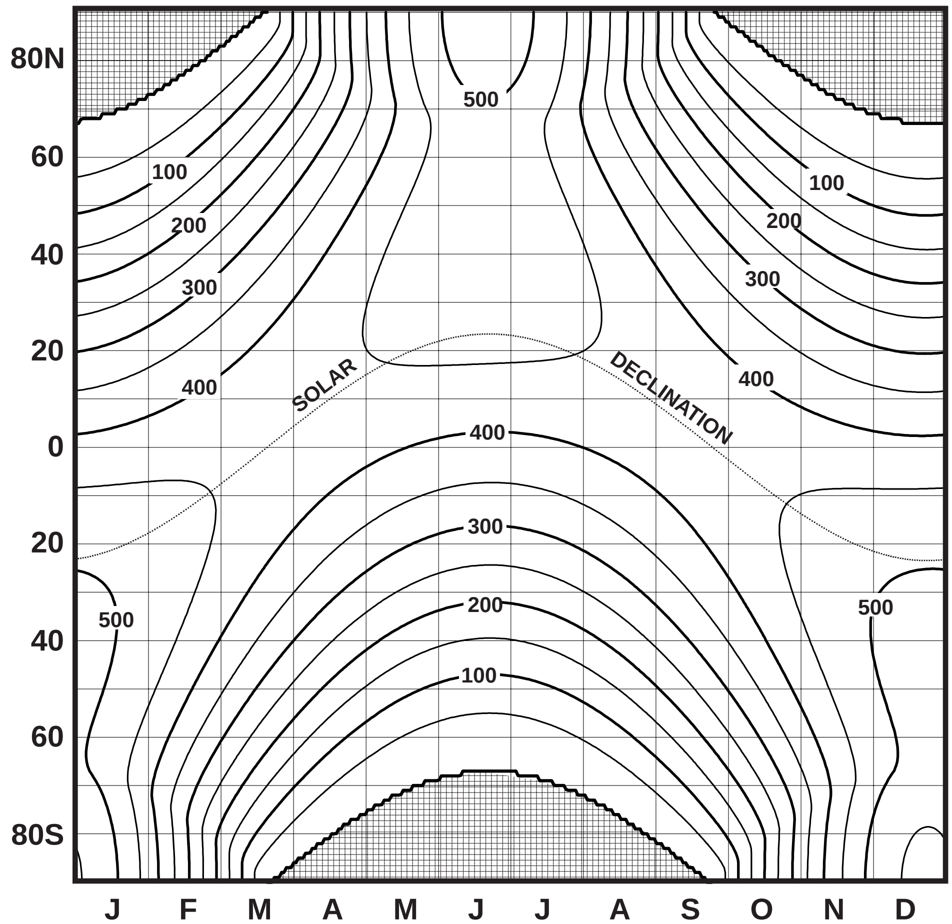

Figure 2.2: The seasonal variation of the zonally (or diurnally) averaged insolation at the top the atmosphere. The units are W m-2.

Near the poles, the length-of-day effect dominates, so that the insolation begins to increase again. That is why there is a minimum of the insolation in the mid-latitudes of the summer hemisphere.

As discussed later, seasonal and, to a lesser extent, diurnal cycles are clearly evident in the circulation patterns. Around the time of the solstices, no insolation at all occurs near the winter pole ("polar night"), but at the same time, near the summer pole, the daily mean insolation is very strong despite low sun angles, simply because the sun never sets ("polar day"). As is well known, these effects arise from the sun - Earth geometry. In addition, the distance from the sun to the Earth varies with time of year, resulting in a few percent more globally averaged insolation in January than in July. The month of maximum insolation varies over geologic time. According to the astronomical theory of the ice ages, extensive glaciation is favored when the minimum insolation occurs during the Northern Hemisphere winter, because the Northern Hemisphere contains about twice as much land as the Southern Hemisphere (e.g., Crowley and North, 1991).

Of course, the infrared radiation emitted by the Earth is very nearly equal to the solar radiation absorbed by the Earth, i.e. it is about 240 W m^{-2}. This near balance between absorbed solar radiation and emitted terrestrial radiation has been directly confirmed by analysis of satellite data. The balance is observed to hold within a few Watts per square meter, which is the within the uncertainty of the measurements.

Fig. 2.3 shows aspects of the Earth's radiation budget as observed in the Earth Radiation Budget Experiment (ERBE; Barkstrom et al., 1989). The zonally averaged incident (i.e. incoming) solar radiation at the top of the atmosphere varies seasonally in response to the Earth's motion around the sun. The zonally averaged albedo, which is the fraction of the zonally averaged incident radiation that is reflected back to space, is highest near the poles, due to snow and ice as well as cloudiness, but it tends to have a weak secondary maximum in the tropics, again associated with high cloudiness there. The zonally averaged terrestrial radiation at the top of the atmosphere, also called the outgoing longwave radiation or OLR, has its maxima in the subtropics. It is relatively small over the cold poles, but it also has a minimum in the warm tropics, due to the trapping of terrestrial radiation by the cold, high tropical clouds, and by water vapor

The net radiation at the top of the atmosphere, which is the difference between the absorbed solar radiation and the OLR, is positive in the tropics and negative in higher latitudes. This implies that energy is transported poleward somehow, inside the system. A portion of this energy is transported by the atmosphere, and the rest is transported by the oceans.

Considering the energy balance of the atmosphere-ocean system, the variation with latitude of the long-term average net radiation at the top of the atmosphere implies energy transports inside the system. These transports are produced by the circulations of both the atmosphere and the oceans, and we can regard the general circulations of the atmosphere and oceans as a “response” to this pattern of net radiation. An important point, however, is that the distributions of the albedo and the outgoing longwave radiation are determined in part by the motion field. It is thus a drastic oversimplification to regard these fields as simple forcing functions; they are bound up with the circulation itself.

Consider the energy budget of a column which extends from the center of the Earth to the “top of the atmosphere”:

$$\frac{\partial E}{\partial t} = N - \nabla \bullet \mathbf{G} . \tag{2.2}$$

Here E is the energy stored in the column; t is time; N is the net radiation input at the top of the atmosphere, with dimensions of energy per unit time per unit area (e.g. W m^{-2}); and the energy transport, $\mathbf{G}$, is a vector with both latitudinal and longitudinal components, and dimensions of energy per unit length per unit time. Later in the course we will discuss how $\mathbf{G}$ can be computed from measurements, but for now we just recognize that it represents the movement of energy due to both the winds and the ocean currents.

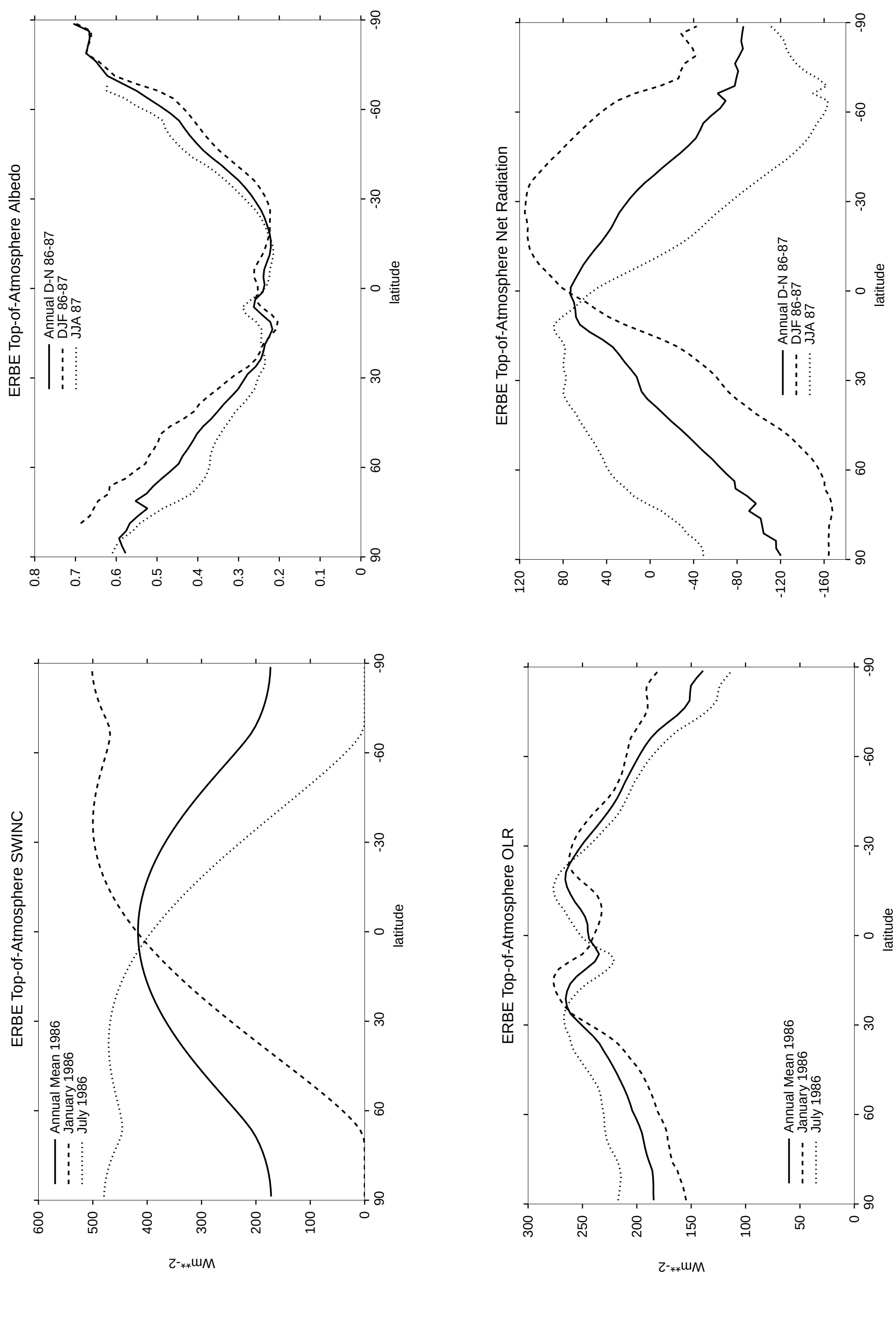

Figure 2.3: The zonally averaged incident solar radiation, albedo, outgoing longwave radiation, and net radiation at the top of the atmosphere, as observed in the Earth Radiation Budget Experiment (Barkstrom 1989).

Suppose that we average (2.2) over a time interval Δt:

$$\frac{E(t+\Delta t)-E(t)}{\Delta t} = \overline{N} - \nabla \bullet \overline{\mathbf{G}}. \tag{2.3}$$

Here the overbar represents the time average. Because the Earth is close to energy balance, $E(t+\Delta t)$ and $E(t)$ cannot be wildly different from each other; this means that the numerator on the left-hand side of (2.3) is bounded in a finite range, regardless of how large Δt is. Therefore, as Δt increases, the left-hand side of (2.3) decreases in absolute value, and eventually becomes negligible compared to the individual terms on the right-hand side. This shows that if the time-averaging interval is long enough then energy storage inside the atmosphere-ocean/land surface at particular locations can be neglected; the minimum time required for this average would be one year, but ideally the average should be taken over many years. In such a time average, the net radiation across the top of the column must be balanced by transports inside; this can be written as

$$\nabla \bullet \overline{\mathbf{G}} = \overline{N}. \tag{2.4}$$

The global mean of $\nabla \bullet \mathbf{G}$ must be zero, at each instant, because the global mean of the divergence of any vector is zero (see Problem 1 at the end of this chapter). This means that the only way that (2.4) can be satisfied everywhere is if the global mean of $N = 0$, i.e. if the Earth is in energy balance.

Now break $\mathbf{G}$ into its zonal and meridional components, i.e.

$$\mathbf{G} = G_\lambda \mathbf{e_\lambda} + G_\varphi \mathbf{e_\varphi}. \tag{2.5}$$

Here $\mathbf{e_\lambda}$ and $\mathbf{e_\varphi}$ are unit vectors pointing towards the east and north, respectively. We expand the divergence operator in spherical coordinates (see Appendix) as follows:

$$\nabla \bullet \mathbf{G} = \frac{1}{a\cos\varphi}\frac{\partial G_\lambda}{\partial \lambda} + \frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}(G_\varphi \cos\varphi). \tag{2.6}$$

Here a is the radius of the Earth. Multiply (2.6) by $a\cos\varphi$ and integrate with respect to longitude, around a latitude circle, to obtain

$$\int_0^{2\pi} (\nabla \bullet \overline{\mathbf{G}}) a\cos\varphi \, d\lambda = \frac{\partial}{\partial\varphi}\int_0^{2\pi} \overline{G_\varphi}\cos\varphi \, d\lambda = \int_0^{2\pi} \overline{N} a\cos\varphi \, d\lambda. \tag{2.7}$$

Note that the zonal derivative has dropped out as a result of the integration. Further integration of (2.7) with respect to latitude, from the South Pole ($\varphi = -\pi/2$) to an arbitrary latitude φ, gives

$$\bar{T}(\varphi) - \bar{T}(-\pi/2) = \int_{-\frac{\pi}{2}}^{\varphi} \int_{0}^{2\pi} \bar{N} a^2 \cos\varphi' d\lambda d\varphi' , \tag{2.8}$$

where $T(\varphi) \equiv \int_0^{2\pi} G_\varphi a \cos\varphi \, d\lambda$. It should be clear that $T(-\pi/2) = T(\pi/2) = 0$. If this were not true, a finite amount of energy per unit time would be flowing into or out of a "point" of zero mass. The right-hand side of (2.8) is just the area integral of N, over the portion of the Earth extending from the South Pole to latitude φ. The dimensions of $T(\varphi)$ are energy per unit time, e.g., Watts. When φ, the upper limit of meridional integration in (2.8), is set to $\pi/2$, the right-hand side of (2.8) simply reduces to the global mean of N, and the left-hand side reduces to zero.

Fig. 2.4 shows a plot of $T(\varphi)$, based on measurements collected in the Earth Radiation Budget Experiment (ERBE; Barkstrom et al., 1989). The poleward energy transport in both hemispheres is clearly apparent. The shape of the curve is roughly like $\sin(2\varphi)$.

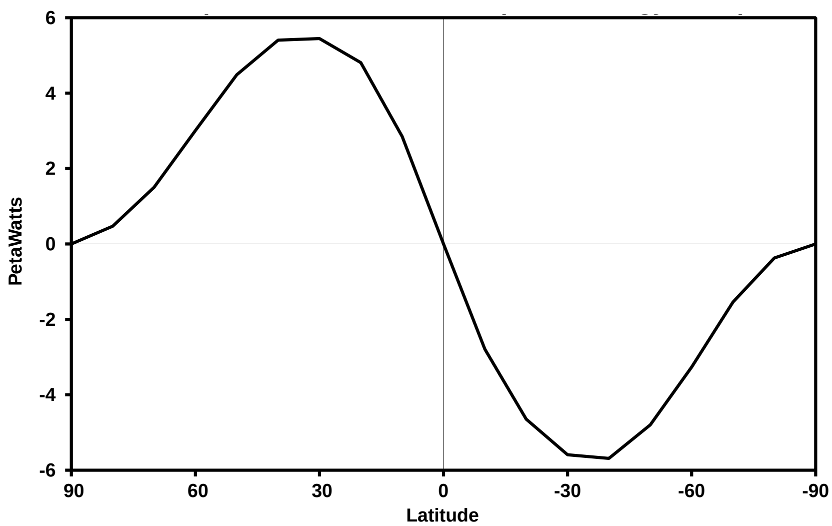

Figure 2.4: The poleward energy transport by the atmosphere and ocean combined, as inferred from the observed annually averaged net radiation at the top of the atmosphere. A Petawatt is 10^{15} W.

We can say that the "job" of the general circulations of the atmosphere and oceans is carry out this meridional energy transport. If the transport of energy from place to place by the atmosphere and oceans could somehow be prevented, then each part of the Earth would have to come into *local* energy balance, by adjusting its temperature, water vapor, and cloudiness so that the outgoing longwave radiation balanced the absorbed solar radiation locally. Such a hypothetical state is referred to as “radiative-convective equilibrium;” modeling studies of radiative-convective equilibrium will be discussed later in this course. Radiative-convective equilibrium would presumably imply much warmer temperatures in the tropics, and much colder temperatures at the poles. The general circulation of the atmosphere and oceans thus has a moderating effect on the global distribution of temperature, tending to warm the higher latitudes and cool the tropics. As we will see, however, these same thermal contrasts represent a source of energy (called “available potential energy”) that makes the global circulations of the atmosphere and oceans possible.

Much further discussion of the observations and theory of energy transports by the atmosphere and oceans is given later in this course.

2.2 Surface boundary conditions

The lower boundary conditions on the general circulation of the atmosphere strongly affect the regional characteristics of the circulation. The most obvious examples are such basic aspects of physical geography as the distribution of land and sea, and the locations of the Earth's mountain ranges shown in Fig. 2.5. These are in part “mechanical” boundary conditions that are independent of season. Note, however, that the land-sea distribution and the locations of “permanent” (or, more accurately, non-seasonal) land ice (e.g., Antarctica and Greenland) strongly affect the surface albedo.

Orography can also provide a thermal forcing, in that the surface of a mountain or an elevated plateau can have a temperature quite different from that of the surrounding air at the same height. For instance, during the northern summer the Tibetan plateau produces a “warm spot” in the middle troposphere, and this represents an important aspect of the thermal forcing that produces the Indian summer monsoon.

“Surface roughness” is another example of a lower boundary condition that is at least partially mechanical in nature. The ocean is relatively smooth, depending on the wind speed, and presents little “roughness” to stimulate momentum exchange with the atmosphere. The land surface is much rougher than the ocean.

One of the most important properties of the Earth’s surface is that roughly 70% of it is permanently wet, and so represents a huge source of moisture. The heat capacity of the ocean is enormous, so that for some purposes (and on sufficiently short time scales, e.g. a few days or weeks) the sea surface temperature (SST) can be considered to be a “fixed” lower boundary condition on the atmosphere. The SST is in this sense the simplest example of a thermal lower boundary condition. The geographical distribution of SST fluctuates seasonally, and shows significant variations with both longitude and latitude, as shown in Fig. 2.6. Note the warm currents off the east coasts of North America and Asia, and the cold currents off all west coasts. The warm SSTs, at a particular latitude, are generally associated with poleward currents; the two best known of these are the Gulf Stream and the Kuroshio. The colder SSTs are generally associated with either equatorward flow (as for example in the case of the California current) or with upwelling (again, in the region of the California current, and also along the Equator in the eastern Pacific). As discussed later, the pattern of upwelling is very closely related to the

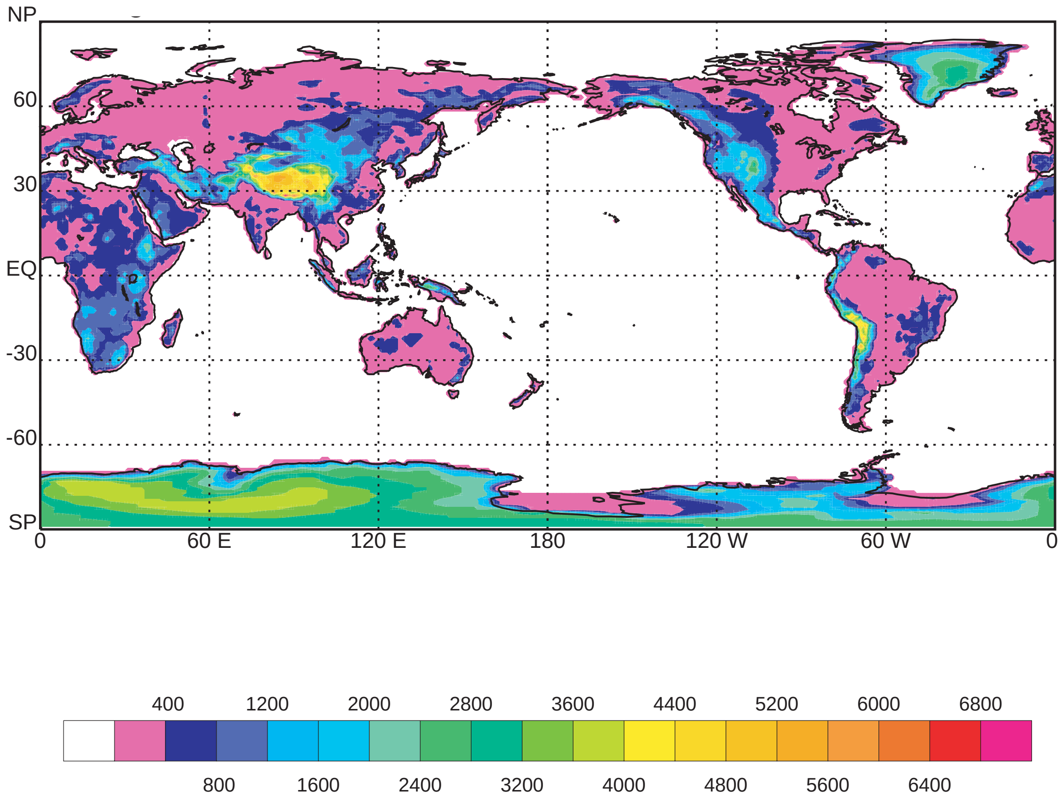

Figure 2.5: The Earth's orography, averaged onto a mesh 1 degree of longitude "wide" by 1 degree of latitude "high."

low-level winds, and at the same time the low-level winds are strongly tied to the spatial distribution of the SST.

The seasonal change of the SST is largest in the Northern Hemisphere, particularly on the western sides of the ocean basins. Note that the seasonal forcing is capable of changing the SSTs by tens of degrees in some middle and high latitude locations. The depth to which this seasonal change penetrates is of course variable, but is typically on the order of 100 m. Of course, the temperature of the water at great depth undergoes virtually no seasonal change.

In the study of the atmospheric general circulation we often consider the spatial and seasonal distribution of the SST to be "given," but of course in reality it is determined in part by what the atmosphere is doing, or rather what the atmosphere has been doing over time. For example, the distribution of cloudiness strongly affects the flow of solar radiation into the upper ocean, and over time this can tend to reduce the SST where clouds are prevalent and the solar insolation at the top of the atmosphere is strong, relative to what the SST would be if the cloudiness were somehow prevented from occurring. The role of clouds in determining the variability of the SST is a major complication hindering our understanding of the atmosphere and ocean as a coupled system.

The distribution of sea ice (Fig. 2.7) also acts as a thermal lower boundary

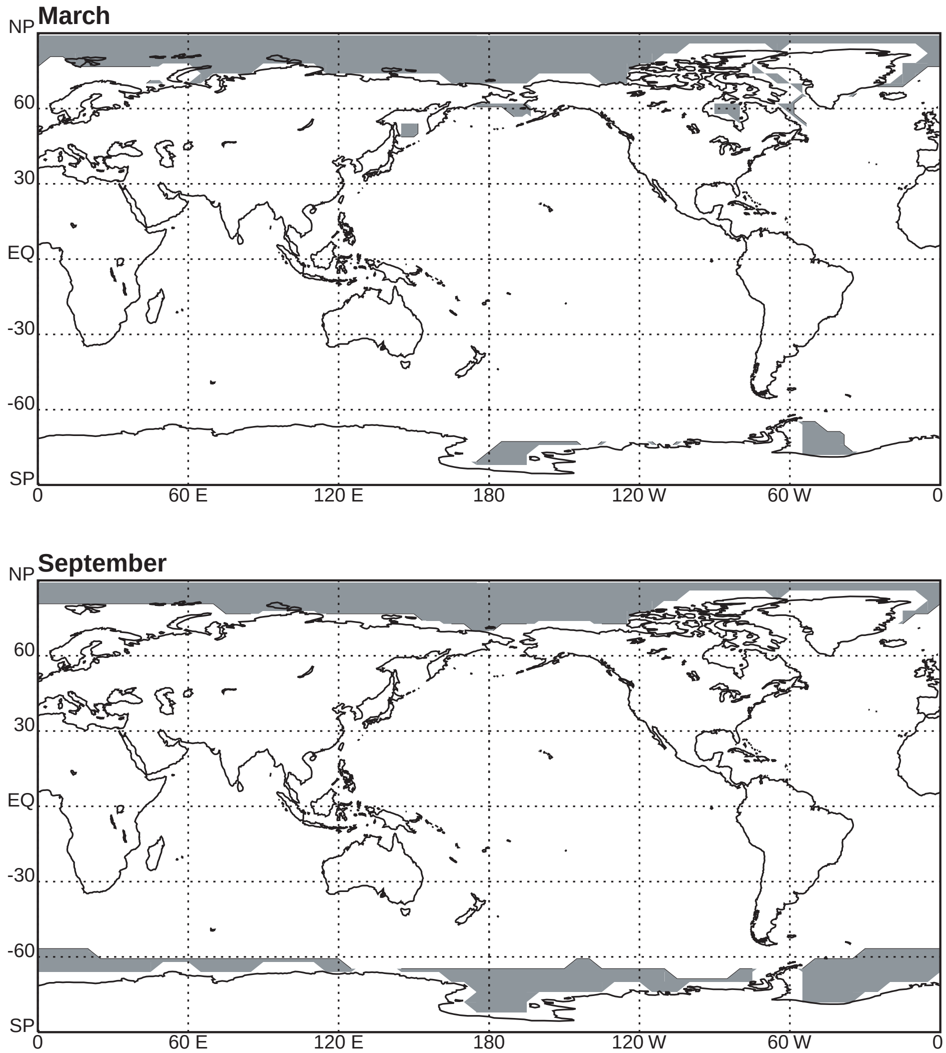

Figure 2.7: The distributions of sea ice (shown in grey) for March and September. The data represent averages for grid cells 5° of longitude wide and 4° of

condition. There are strong seasonal changes in ice cover in the Southern Hemisphere, but not in the Northern Hemisphere. In addition to the obvious strong effect of sea ice on the surface albedo, the ice also acts as an insulator that separates the relatively warm ocean water from the air. Because sea ice is such a good insulator, its upper boundary can be much colder than the water beneath. Sea ice is also very smooth, so that little surface drag occurs for a given wind speed. The Arctic ocean is ice-covered all year, while the North Atlantic and the Southern Oceans experience seasonal melting. Of course, the thickness of the ice also varies both geographically and seasonally. In addition, several percent of open water typically occurs, especially when the ice is thin. This open water is often in the form of cracks called "leads." The open water in leads is often much warmer

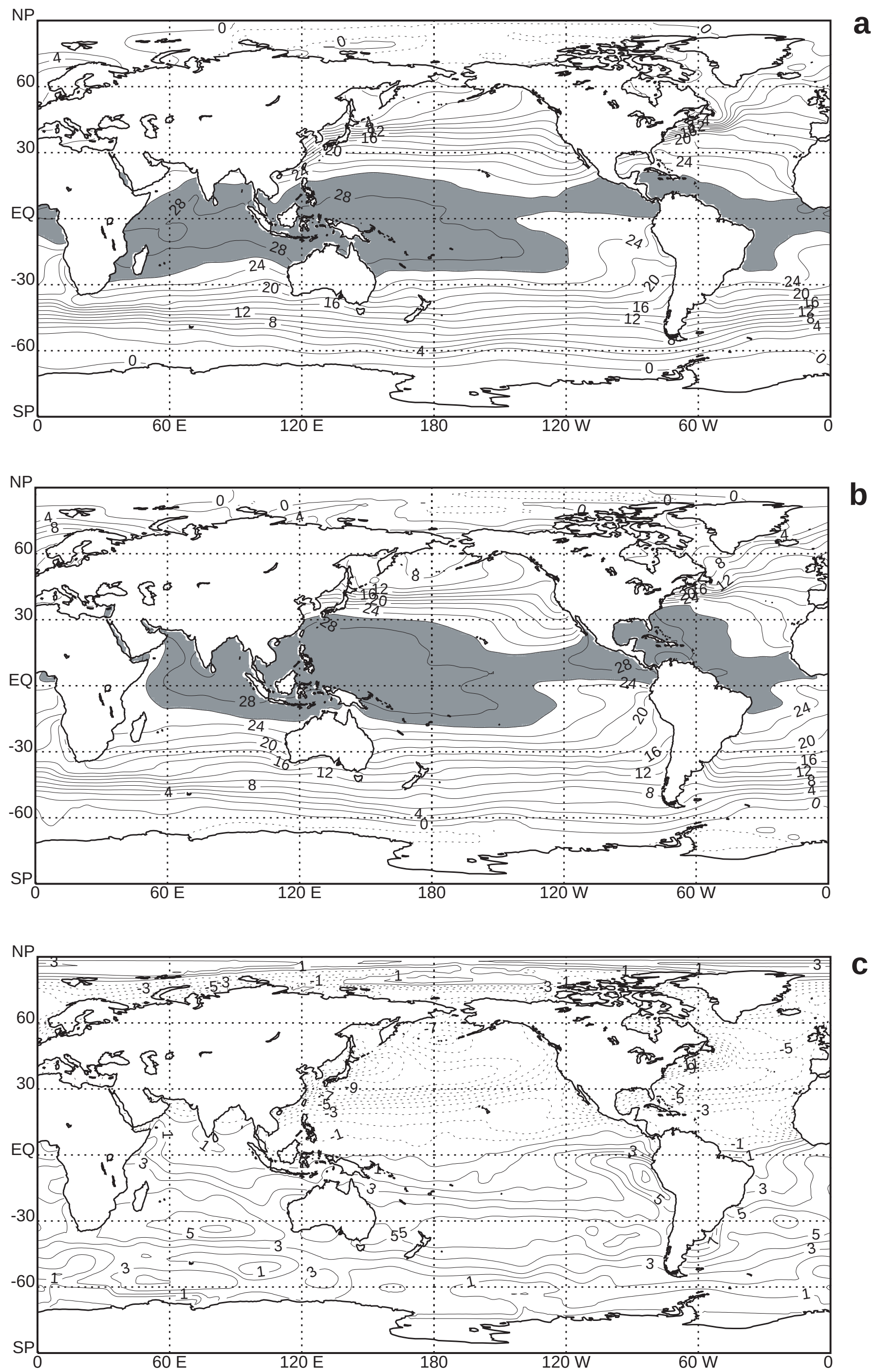

Figure 2.6: a) Sea surface temperature distribution for January. The contour interval is 2 K, and values greater than 26 °C are shaded. b) Same for July. c) The sea surface temperature for March, minus the sea surface temperature for September. The contour interval is 2 K. The zero line is heavy, and negative values are indicated by dashed contours.

than the ice nearby, especially in winter. Under such conditions, the large-scale average sensible and latent heat fluxes are typically dominated by the contributions from the leads, even though leads may cover only a few percent of the area. Snow that falls on the sea ice insulates it and protects it from the effects of the sun, helping to prevent the ice from melting. For a perspective on atmosphere-ocean-sea ice interactions, see Randall et al. (1998).

The pattern of vegetation on the land surface affects the atmosphere in very complicated ways. A simplified summary of the observed distribution of vegetation on the land surface is given in Fig. 2.8. Obviously, the type, density, and even the health of

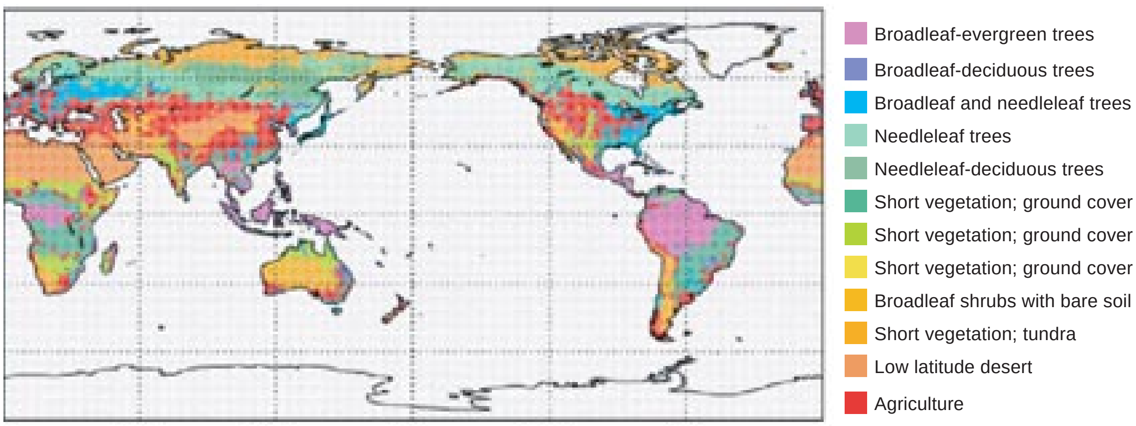

Figure 2.8: A simplified depiction of the distribution of vegetation on the land surface. The resolution is one degree. "Permanent" land ice is shown in white.

the land-surface vegetation can affect the surface albedo and surface roughness. These characteristics of the vegetation vary with season, especially in middle latitudes. They can also vary interannually. The degree to which the plants allow moisture to transpire from leaves into the atmosphere strongly regulates the surface fluxes of sensible and latent heat; strong transpiration cools the surface and reduces the sensible heat flux. Sellers et al. (1997) provide an introductory overview.

The geographical and seasonal variations of the surface albedo are largely determined by the distribution of vegetation, but of course they also depend on snow cover.

Permanent land ice, also shown in Fig. 2.8, is mainly confined to Antarctica and Greenland, in the present climate. The distribution of land ice can vary dramatically on time scales of thousands of years and longer (e.g. Imbrie and Imbrie, 1979).

Some aspects of the atmospheric general circulation can be regarded as more or less direct responses to the various boundary "forcings" mentioned above. Examples include the equator-to-pole energy flux, planetary waves produced by flow over mountains, and monsoons that are strongly tied to the land-sea distribution and the seasonally varying insolation. Of course, there are many additional time-dependent features of the circulation that are less directly tied to the boundary conditions, but instead arise from the internal dynamics of the atmosphere. Examples include baroclinic waves and organized moist convection.

2.3 Energy and moisture budgets of the surface and atmosphere

The planetary radiation budget has already been briefly discussed. We now consider the energy and moisture budgets of the Earth's surface and the atmosphere. It is a shocking fact that we do not know enough about the globally averaged surface energy budget to do more than sketch rough annual mean values, as shown in Table 2.2 . None of

Table 2.2: Components of the globally and annually averaged surface energy budget. A positive sign means that the surface is warmed.

Absorbed solar (SW)	176 W m^{-2}
Downward infrared (LW↓)	312 W m^{-2}
Upward infrared (LW↑)	-385 W m^{-2}
Net longwave (LW)	-73 W m^{-2}
Net radiation	103 W m^{-2}
Latent heat (LH)	-79 W m^{-2}
Sensible heat (SH)	-24 W m^{-2}

the numbers in the table is known to better than 20% accuracy. Of the 240 W m^{-2} that is absorbed by the Earth-atmosphere system, 176 W m^{-2} is absorbed by the Earth's surface. Thus only about 240 - 176 = 64 W m^{-2} of solar radiation is absorbed by the atmosphere. That is only about 1/4 of the total solar radiation absorbed by the Earth-atmosphere system. It should be noted, however, that the partitioning of the absorbed solar radiation between the atmosphere and the Earth's surface is currently a matter of some controversy.

The surface receives a total (LW↓ + SW; see notation defined in Table 2.2) of 488 W m^{-2}, which is given back in the form of LW↑, LH and SH. By far the largest of these is LW↑. Keep in mind that the oceans can transport energy from one place to another, so that the energy absorbed by the oceans is not necessarily given back in the same place where it is absorbed. Also, the large heat capacity of the upper ocean allows energy storage on seasonal time scales. In contrast, the continents cannot transport energy internally at a significant rate, and their limited heat capacity forces near energy balance, everywhere, on time scales longer than a few days (at most).

Note that *the net radiative heating of the surface, which amounts to 103 W* m^{-2}*, is balanced primarily by evaporative cooling of the surface at the rate of 79 W* m^{-2}. As discussed below, moisture is of comparable importance in the energy budget of the atmosphere.

The globally averaged energy budget of the atmosphere is shown in Table 2.3.

Table 2.3: The globally and annually averaged energy budget of the atmosphere. A positive sign means that the atmatmospghere is warmed.

Absorbed solar radiation (240 - 176)	64 W m^{-2}
Net emitted terrestrial radiation (-240 + 73)	-167 W m^{-2}
Net radiative heating	-103 W m^{-2}
Latent heat input	79 W m^{-2}
Sensible heat input	24 W m^{-2}

Again, most of the numbers in Table 2.2 and Table 2.3 are only rough estimates. One interpretation of Table 2.3 is that the atmosphere sheds energy through infrared radiation at the rate required to balance the various forms of energy input, and the temperature of the atmosphere is that required to allow the necessary infrared emission.

We can also say that the net radiative cooling of the atmosphere, at the rate of -103 W m^{-2}, is primarily balanced by the latent energy source due to surface evaporation. Of course, this latent energy is converted into sensible heat when water vapor condenses. A fraction of the condensed water re-evaporates inside the atmosphere. The net condensation rate within the atmosphere is closely balanced by the rate of precipitation at the Earth's surface; this means that the amount of condensed water in the atmosphere is neither increasing nor decreasing with time. The rate at which evaporation introduces moisture into the atmosphere is balanced by the rate at which precipitation removes it. Keep in mind that these various balances apply in a globally averaged sense, rather than locally in space, and in a time-averaged sense, rather than instantaneously.

The preceding discussion suggests a second interpretation of the atmospheric energy budget: to a first approximation, *the globally averaged precipitation rate is "determined by" the rate at which the atmosphere is cooling radiatively.* Of course, this does not mean that the geographical and temporal distributions of precipitation are determined by the corresponding distribution of radiative cooling; in fact, the *local* rate of precipitation tends to be *negatively* correlated with the local atmospheric radiative cooling, and is controlled mainly by dynamical processes.

The net radiative cooling of the atmosphere is strongly affected by the high, cold cirrus clouds, many of which are formed within precipitating cloud systems. The cirrus clouds absorb the infrared radiation emitted by the warm atmosphere and surface below; the cirrus themselves emit much more weakly because they are very cold. This means that the cirrus effectively trap infrared radiation inside the atmosphere. For this reason, as the cirrus cloud amount increases, the radiative cooling of the atmosphere decreases.

Consider together the following points which have been made in the last few paragraphs:

- The radiative cooling of the atmosphere is primarily balanced by latent heat release in precipitating cloud systems.
- Precipitating weather systems produce cirrus clouds.
- Cirrus clouds tend to reduce the radiative cooling of the atmosphere.

Taken together, these points suggest a negative feedback loop which tends to regulate the strength of the hydrologic cycle. To see how this works, suppose that we have an equilibrium in which atmospheric radiative cooling and latent heat release are in balance. Suppose that we perturb the equilibrium by increasing the speed of the hydrologic cycle, including the rate of latent heat release. The same perturbation will increase the rate of cirrus cloud production, which will reduce the rate at which the atmosphere is radiatively cooled. In order to restore atmospheric energy balance, it will be necessary for the hydrologic cycle to slow down, i.e., the initial perturbation will be damped. Fowler and Randall (1994) give further discussion.

The “effective altitude” for infrared emission by the Earth-atmosphere system is near 5 km above sea-level. This simply means that the outgoing longwave radiation at the top of the atmosphere is equivalent to that from a black body whose temperature is that of the atmosphere near the 5 km level. Roughly speaking, then, atmospheric motions must carry energy upward from the surface through the first 5 km of the atmosphere, and infrared emission carries the energy the rest of the way out to space. This upward energy transport by circulating air occurs on both small scales, notably in boundary-layer turbulence and cumulus convection, and also on large scales, notably through midlatitude baroclinic eddies and the tropical Hadley circulation. In short, *the atmospheric circulation carries energy upward as well as poleward.* As discussed later, convective clouds play a particularly important role in the upward energy transport.

We now examine in more detail the fluxes of various quantities at the Earth’s surface. In addition to the surface solar and terrestrial radiation, we must also consider the turbulent fluxes of momentum, sensible heat, and latent heat. In principle, we should also consider the fluxes of various chemical species, but this important aspect of the climate system is neglected here.

The seasonal variations of the surface shortwave and longwave radiation at a particular station are illustrated in Fig. 2.9, which shows the variations of the upward and downward shortwave (SW) and longwave (LW) near-surface radiation at the Boulder Atmospheric Observatory (BAO) tower, near Boulder Colorado. (The tower is located about 1 km west of Interstate 25, and slightly south of Colorado Route 52.) The data cover the three years 1986 - 89. The seasonal cycle is clearly evident. High frequency fluctuations are primarily due to cloudiness. Note that the upward solar radiation has maxima during the winter. These are associated with the increased albedo of the ground following snow storms.

Data like those shown in Fig. 2.9 are available for only a few stations around the world. Most of our ideas about the global pattern of surface radiation are based on various estimates, which have been carefully worked out but are subject to significant errors.

Fig. 2.10 shows the meridional distribution of the solar radiation absorbed by the

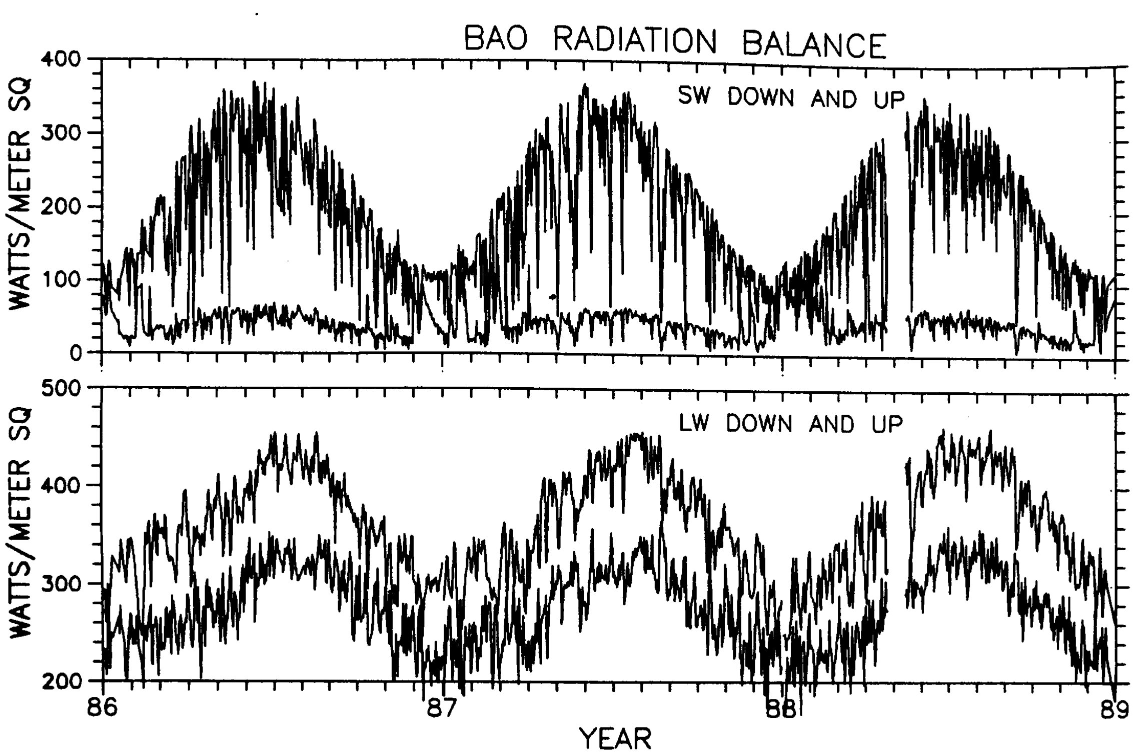

Figure 2.9: Observations of the upward and downward solar (SW) and terrestrial (LW) radiation at the Boulder Atmospheric Observatory (BAO) tower near Boulder Colorado, for the years 1986 - 1988. Figure provided by Ellsworth Dutton.

Earth's surface, as a function of latitude, and for January, July, and the annual mean. The seasonal changes associated with the movement of the Earth in its orbit are clearly visible. These have already been discussed, earlier in this chapter. There are additional features that require some explanation. For example, near 50 °N in July there is a minimum of the surface absorbed solar radiation. This is associated with cloudiness in those latitudes, and indicates that the clouds are having a major impact on the energy budget of the ocean in those latitudes. A weaker tropical minimum occurs for the same reason. The annual mean curve is fairly symmetrical about the Equator, but shows a minimum near 10 °N associated with the ITCZ. Note also that in the annual mean the southern high latitudes absorb less than the northern high latitudes.

The zonally averaged net surface longwave energy flux is shown in Fig. 2.11. Here there are some real surprises. Although the ocean temperatures are warmer in summer than in winter, the net longwave cooling of the surface is actually stronger in winter than in summer! This occurs despite the fact that the surface emission is essentially proportional to T^4, and so must be considerably larger in summer than in winter. The explanation is that the downward radiation from the atmosphere to the surface increases from winter to summer due to both the warming of the air and the increase in the atmospheric emissivity due to seasonally increased water vapor content and also seasonal changes in cloudiness. This increase in the downward component is so strong that it overwhelms the increase in the upward component, giving a net decrease in surface infrared cooling from winter to summer. In January, the strongest cooling occurs over Antarctica, and in the subtropics of the winter hemisphere. The weakest cooling

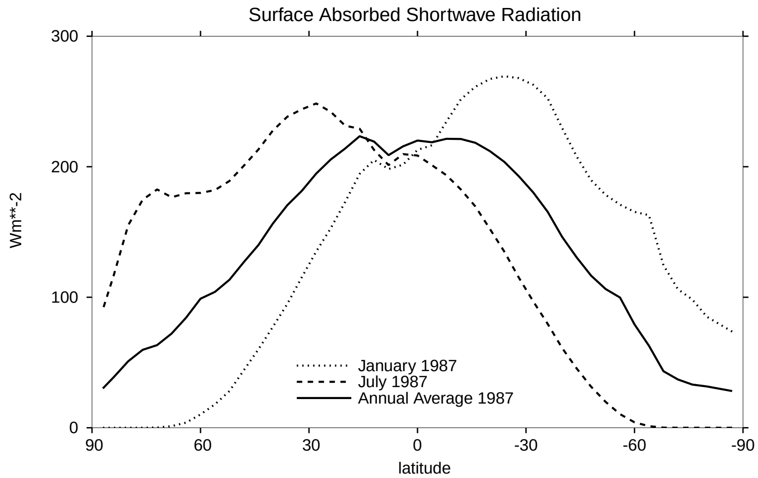

Figure 2.10: The zonally averaged (land and ocean) net solar radiation absorbed by the Earth's surface. These estimates are uncertain by 5% or so. Based on Darnell et al. (1988, 1992). Note: These data are *not* true observations, although they are based on observations.

occurs in cloudy regions, e.g. over the Southern Oceans and in the storm tracks of both hemispheres.

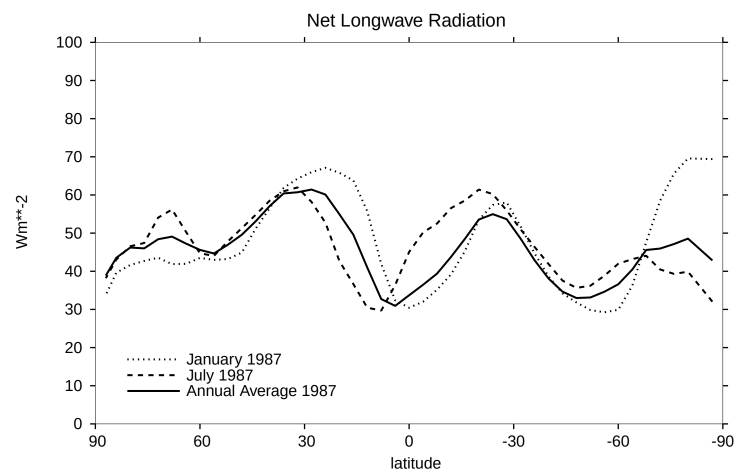

Figure 2.11: The zonally averaged (land and ocean) net infrared cooling of the Earth's surface. These estimates have large but difficult to quantify uncertainties. Based on Gupta (1989) and Gupta et al. (1992). Note: These data are *not* true observations, although they are based on observations.

Fig. 2.12 shows the zonally averaged net surface radiation (solar and terrestrial combined). High latitudes experience net radiative cooling of the surface in winter, as would be expected. The annual mean net radiation into the surface is positive at all latitudes. Clearly the surface must cool by non-radiative means.

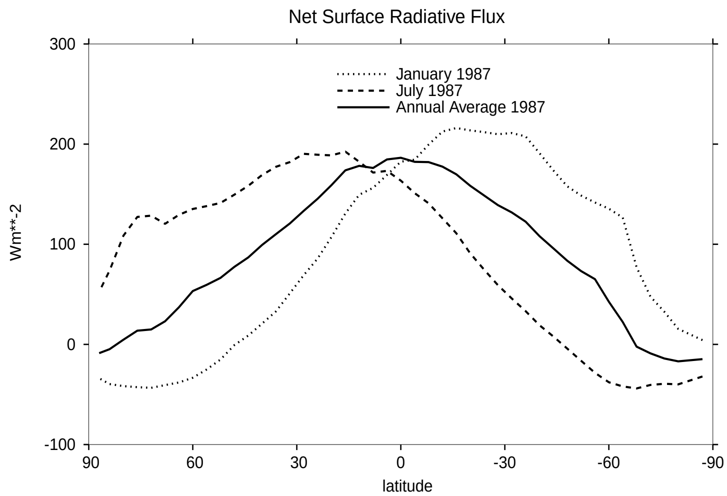

Figure 2.12: **The zonally averaged net surface radiation, obtained by combining the data from Fig. 2.10 and Fig. 2.11. Note: These data are *not* true observations, although they are based on observations.**

Fig. 2.13 shows the zonally averaged net latent heat flux, as reported by Esbensen and Kushnir (1981). Positive values represent a moistening of the atmosphere and a cooling of the surface. Clearly the latent heat flux compensates, to a large extent, for the net radiative heating of the surface shown in the previous figure. Note that the maxima of the latent heat flux occur in the subtropics. Recall that the precipitation maxima occur in the tropics and middle latitudes. This implies that moisture is transported from the subtropics into the tropics, and from the subtropics into middle latitudes.

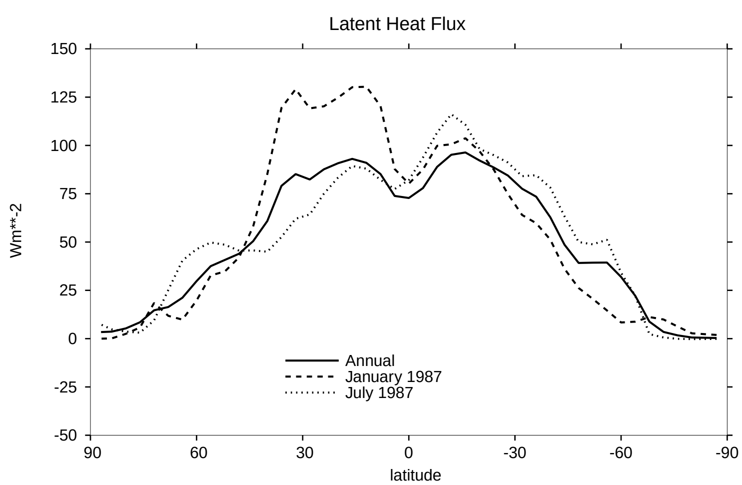

Figure 2.13: **The zonally averaged surface latent heat flux, positive upward, based on ECMWF analyses for 1987. Note: These data are *not* true observations, although they are based on observations.**

Fig. 2.14 shows the corresponding curves for the surface sensible heat flux. Note that the surface sensible heat flux is generally smaller than the surface latent heat flux. Maxima occur in the winter hemisphere, especially in the Northern winter in association with cold-air outbreaks over warm ocean currents at the east coasts of North America and Asia. Local heat flux maxima associated with such cold outbreaks can be on the order of 1000 W m^{-2}, on individual days.

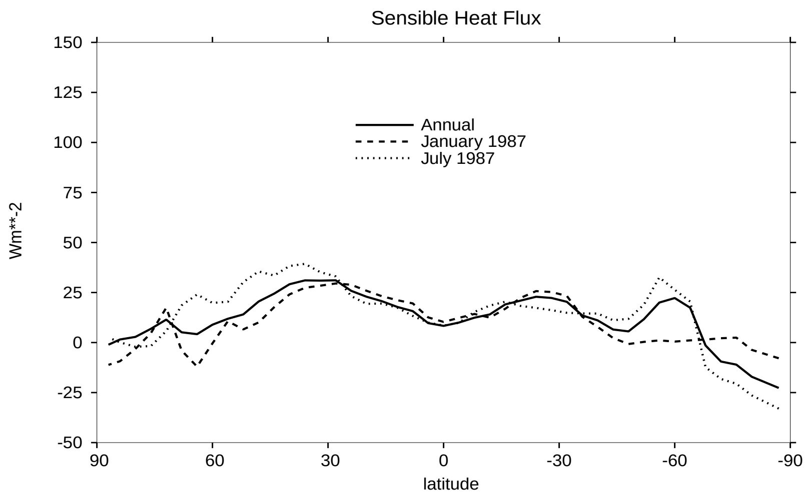

Figure 2.14: The zonally averaged surface sensible heat flux, positive upward, based on ECMWF analyses for 1987. Note: These data are *not* true observations, although they are based on observations.

2.4 Summary

This chapter provides a summary of the energy fluxes at the top of the atmosphere, at the Earth's surface, and across the atmosphere. The meridional structure of the net radiation at the top of the atmosphere implies transports by the ocean-atmosphere system. The meridional structure of the net surface energy flux implies energy transports by the ocean. The meridional structure of the net flow of energy into the atmosphere, across its upper and lower boundaries, implies a net transport of energy by the atmosphere. In later chapters, we will discuss the nature of these transports in more detail; for now we simply note that they must occur.

We have also discussed the lower boundary conditions on the atmosphere, which are provided by the Earth's surface. These involve many complex factors including the distributions of continents and oceans, the arrangements of the mountains, and the distribution of vegetation.

We also discussed the moisture budget of the Earth's surface and the atmosphere. Among the most important points to emerge from this analysis is that the net radiative heating of the Earth's surface is balanced mainly by evaporative cooling, and the net radiative cooling of the atmosphere is balanced mainly by latent heat release. We have also noted the important effects of water vapor and clouds on the Earth's radiation budget, and the effects of ice, snow, and the land-sea distribution on the radiative properties of the Earth's surface. These facts make it clear that the hydrologic cycle plays a very central role in the general circulation of the atmosphere, and in the Earth's energy balance. This point will be made again later in a variety of ways.

We must also discuss exchanges of momentum between the atmosphere and the Earth's surface; this is postponed until later.

Problems

1. a) Prove that, for any vector $\mathbf{Q}$,

$$\int_s \nabla \bullet \mathbf{Q} \, dS = 0, \tag{2.9}$$

where the integral is taken over a closed surface.We assume that $\mathbf{Q}$ is everywhere tangent to the surface, i.e. it "lies in" the surface. Note: This shows that the globally averaged divergence of any "horizontal" vector field is zero.

b) Also prove that

$$\int_s \mathbf{k} \bullet (\nabla \times \mathbf{Q}) \, dS = 0 \tag{2.10}$$

where the integral is take over a closed surface. Here $\mathbf{k}$ is a unit vector everywhere perpendicular to the surface. Note: This shows that the globally averaged vorticity is zero.

2. Suppose that at 40° N the northward energy transport shown in Fig. 2.4 is entirely due to the atmosphere, and is produced by the combination of a northward wind around half of the latitude circle, and a compensating southward wind in the other half of the latitude circle, with uniform speeds of 5 m s^{-1} each. For simplicity, suppose that these currents fill the entire depth of the atmosphere, and that the surface pressure is a uniform 1000 mb. Assume that the temperature is zonally uniform within each current. Compute the implied temperature difference between the northward and southward flows.

3. a) Suppose that 1 W m^{-2} is supplied to a column of water 100 m deep. How much time is needed to increase the temperature of the water by 1 K?

b) Estimate the heat capacity of the entire global ocean in J K^{-1}. If all of the solar radiation incident on the top of the atmosphere were used to warm the ocean uniformly, how long would it take to increase the temperature of the entire ocean by 1 K?