2. What makes it go?

The Earth’s radiation budget: An “upper boundary condition” on the global circulation

Radiation is (almost) the only mechanism by which the Earth can exchange energy with the rest of the Universe. The most important upper boundary condition on the global circulation of the atmosphere is the incident solar radiation, also called the insolation. The insolation varies with geographical location, and with time. It is determined by the energy output of the Sun, the spherical geometry of the Earth itself, and the geometry of the Earth’s orbit (see Fig. 2.1), which can be described in terms of the obliquity (the angle that the Earth’s axis of rotation makes with the plane of the Earth’s orbit around the sun), the eccentricity (a measure of the degree to with

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 https://dept.astro.lsa.umich.edu/ugactivities/Labs/seasons/SeasonsIntro.tropics.html.
the shape of the Earth’s orbit differs from a perfect circle), and the dates of the equinoxes. These
all vary over geologic time (e.g., Crowley and North, 1991).

The solar energy flux at the mean radius of the Earth's orbit is about 1365 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 1365 W m$^{-2}$ is equivalent to 1.365 GW km$^{-2}$, which is the
energy output of a large power plant for each square kilometer of area normal to the solar beam.

<table>
<thead>
<tr>
<th>Incident solar radiation</th>
<th>341 W m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Absorbed solar radiation</td>
<td>239 W m$^{-2}$</td>
</tr>
<tr>
<td>Planetary albedo</td>
<td>0.30</td>
</tr>
<tr>
<td>Outgoing longwave radiation</td>
<td>239 W m$^{-2}$</td>
</tr>
</tbody>
</table>

Table 2.1 Summary of the annually averaged top-of-the-atmosphere radiation budget, after Trenberth et al. (2009).

The globally averaged top-of-the-atmosphere radiation budget is summarized in Table 2.1.
The numbers given in the table are now known to three significant digits. The Earth’s albedo is
near 0.30, independent of season; this number has been known to better than 10% accuracy only
since the 1970s. The energy absorbed by the Earth is

$$S_{\text{abs}} = S \left( \frac{\pi a^2}{4 \pi a^2} \right) (1 - \alpha)$$

$$= \frac{1}{4} S (1 - \alpha).$$

(1)

Here $S_{\text{abs}}$ is the globally averaged absorbed solar energy per unit area (given in Table 2.1), $S$ is
the globally averaged insolation, $a$ is the radius of the Earth, and $\alpha$ is the planetary albedo. The
global average used to compute $S_{\text{abs}}$ includes the zeroes on the night-side of the Earth. That is
why $S_{\text{abs}}$ is multiplied by $\pi a^2$, the area of the absorbing disk, and divided by $4 \pi a^2$, the area of
the sphere.

As a matter of common experience, the insolation varies both diurnally and seasonally. At a
given moment, it 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 (i.e., the latitude immediately “under” the Sun) 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. On the other hand, the length of day increases at high latitudes in summer, and this tends
to make the daily-mean insolation increase. Near the poles, the length-of-day effect dominates, so that at high latitudes in summer the insolation actually increases towards the pole. That is why there is a minimum of the insolation about 23° away from the pole in the summer hemisphere, as shown in the figure.

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 (the 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 (the polar day). As is well known, these effects arise from the sun-Earth geometry shown in Fig. 2.1. In addition, the distance from the sun to the Earth varies slightly with time of year, resulting in a few percent more globally averaged insolation in January than in July, in the current epoch. The month of maximum insolation varies over geologic time. According to the widely accepted astronomical theory of the ice ages, extensive glaciation is favored when the minimum insolation occurs during the Northern Hemisphere summer, because the Northern Hemisphere contains about twice as much land as the Southern Hemisphere (e.g., Crowley and North, 1991).

Figure 2.2: The seasonal variation of the zonally (or diurnally) averaged insolation at the top of the atmosphere. The units are W m$^{-2}$.
The Earth emits about as much infrared radiation as required to balance the absorbed solar radiation; both energy flows are at the rate of about 240 W m$^{-2}$. This near balance has been directly confirmed by analysis of satellite data. The balance is observed to hold within a few Watts per square meter, which is comparable to the uncertainty of the measurements. The actual annual-mean imbalance is thought to be about 0.5 W m$^{-2}$ (Loeb et al., 2012; Trenberth et al., 2014).

Fig. 2.3 shows aspects of the Earth's radiation budget as observed from satellites (Wielicki et al., 1998). 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 cloudiness as well as snow and ice. It tends to have a weak secondary maximum in the tropics, 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.
Figure 2.3: The zonally averaged a) incident solar radiation, b) absorbed solar radiation at the top of the atmosphere, c) albedo, d) outgoing longwave radiation, and e) net radiation at the top of the atmosphere, as observed using satellites in a project called “Clouds and the Earth’s Radiant Energy System” (CERES; Wielicki et al., 1998).
Meridional energy transports by the atmosphere-ocean system

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 global 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 strongly influenced by the circulation itself.

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

\[
\frac{\partial E}{\partial t} = N_\infty - \nabla \cdot G_\infty.
\]  

(2)

Here \( E \) is the energy per unit area stored in the column; \( t \) is time; \( N_\infty \) is the net downward flux of energy at the top of the atmosphere, which is entirely due to radiation. \( N_\infty \) has dimensions of energy per unit time per unit area (e.g., W m\(^{-2}\)). The energy transport, \( -\nabla \cdot G_\infty \), represents the movement of energy, in the zonal and meridional directions, due to both the winds and the ocean currents, and \( G_\infty \) is a vector with both zonal and meridional components, and dimensions of energy per unit length per unit time (e.g., W m\(^{-1}\)). The subscript \( \infty \) on \( G_\infty \) means that it includes all parts of the Earth system, from the center of the Earth out to space.

Suppose that we average (2) over a time interval \( \Delta t \):

\[
\frac{E(t + \Delta t) - E(t)}{\Delta t} = \overline{\left[ N_\infty - \nabla \cdot G_\infty \right]}.
\]  

(3)

Here \( \overline{()} \) represents a time average; this is a notation that we will use throughout the book. 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 (3) is bounded within a finite range, regardless of how large \( \Delta t \) is. Therefore, as \( \Delta t \) increases, the left-hand side of (3) decreases in absolute value, and eventually becomes negligible compared to the individual terms on the right-hand side. The physical meaning is that energy storage inside the Earth system at particular locations can be neglected if the time-averaging interval is long enough; the minimum time required for such an average would be one year, but ideally the average should be taken over many years. When we apply such a time average, the net radiation across the top of the column must be balanced by transports inside; this can be written as
\[ \nabla \cdot G_\infty = N_\infty. \]  

(4)

The global mean of \(-\nabla \cdot G_\infty\) must be exactly zero, not just in a time average, but at each instant. The reason is purely mathematical, rather than physical: the global mean of the divergence of any vector is zero; you are asked to prove this in Problem 1 at the end of this chapter. The only way that (4) can be true is if the global mean of \(N_\infty\) is equal to zero, i.e., if the Earth is in energy balance. To the extent that this is not true, the left-hand side of (3) is not negligible.

We now break \(G_\infty\) into its zonal and meridional components, i.e.

\[ G_\infty = (G_\infty)_\lambda e_\lambda + (G_\infty)_\varphi e_\varphi. \]  

(5)

Here \(e_\lambda\) and \(e_\varphi\) are unit vectors pointing towards the east and north, respectively. The symbols \(\lambda\) and \(\varphi\) represent longitude and latitude, respectively. We expand the divergence operator in spherical coordinates (see the Appendix on “Vectors, Vector Calculus, and Coordinate Systems”) as follows:

\[ \nabla \cdot G_\infty = \frac{1}{a \cos \varphi} \frac{\partial (G_\infty)_\lambda}{\partial \lambda} + \frac{1}{a \cos \varphi} \frac{\partial}{\partial \varphi} \left[(G_\infty)_\varphi \cos \varphi\right]. \]  

(6)

Here \(a\) is the radius of the Earth. Multiply both sides of (6) by \(a \cos \varphi\) and integrate over all longitudes to obtain

\[
\int_0^{2\pi} (\nabla \cdot G^\prime_\infty) a \cos \varphi d\lambda = 2\pi \frac{\partial}{\partial \varphi} \left[(G_\infty)_\varphi \cos \varphi\right] = 2\pi N_\infty a \cos \varphi.
\]  

(7)

Here we use the notation

\[ \overline{()}^\lambda \equiv \frac{1}{2\pi} \int_0^{2\pi} () d\lambda. \]  

(8)
to denote a zonal mean; the *combination* of a time average and a zonal mean can be denoted by either $\langle \cdot \rangle^{\lambda,s}$ or $\langle \cdot \rangle^{s,\lambda}$. In (7), the zonal derivative has dropped out as a result of the integration with respect to longitude. The second line of (7) comes by comparison with (4).

Eq. (7) gives us a way to compute the meridional derivative of the poleward energy transport, i.e., $\frac{\partial}{\partial \phi} \int_0^{2\pi} (G_{\infty})_{\phi} \cos \phi d\lambda$, in terms of $N_{\infty}^{\lambda,s}$. What we really want, however, is a formula for the poleward energy transport itself, rather than its divergence. To get this, we multiply (7) by $a$, and integrate with respect to latitude, from the South Pole ($\phi = -\pi/2$) to an arbitrary latitude $\phi$, to obtain

$$\overline{\Theta}'(\phi) = 2\pi a^2 \int_{-\pi/2}^{\phi} N_{\infty}^{\lambda,s} \cos \phi' d\phi', \tag{9}$$

where we define

$$\Theta(\phi) \equiv 2\pi a \cos \phi (G_{\infty})_{\phi}^{\lambda,s}, \tag{10}$$

and we used the boundary condition

$$\Theta(-\pi/2) = 0. \tag{11}$$

The dimensions of $\Theta(\phi)$ are energy per unit time, e.g., Watts. The boundary condition (10) is *exact*; if it were not true, a finite amount of energy per unit time would be flowing into or out of the South Pole, which is a “point” of zero mass. A similar condition must apply at the North Pole. The right-hand side of (9) is the *area integral* of $N_{\infty}'$ over the “south polar cap” that extends from the South Pole up to latitude $\phi$. When the upper limit of meridional integration in (8), is set to $\pi/2$, the right-hand side of (9) reduces to the global mean of $N_{\infty}'$.

Fig. 2.4 gives a plot of $\overline{\Theta}'(\phi)$, computed by using (9), with values of $N_{\infty}'$ based on satellite data from a project called CERES (Wielicki et al., 1996, 1998). A poleward energy transport is clearly apparent in both hemispheres. The curve of the transport has a pleasingly simple shape,  

---

1 There is a long tradition of using square brackets to denote zonal means. I have decided not to follow the tradition, and hope that others will do the same.
roughly like $\sin(2\varphi)$ . Vonder Haar and Oort (1973) were the first to diagnose $\overline{\Theta}'(\varphi)$ from satellite measurements of the Earth’s radiation budget. The maximum absolute values in middle latitudes of both hemispheres are on the order of 6 PW (a PetaWatt is $10^{15}$ Joules per second). Because this total energy transport by the climate system can be inferred directly from satellite measurements of the Earth’s radiation budget, it is known with relatively good accuracy now.

Fig. 2.4 shows that that $\overline{\Theta}'(\varphi)$ is exactly zero at both poles. We built that in for the South Pole, by using (11), but what makes $\overline{\Theta}'(\varphi)$ miraculously return to zero at the North Pole? The answer is that we have forced it to be zero at the North Pole, by “correcting” the data. The observations show that the global mean of $\overline{N}_\infty'$ is small compared to the local values. It is zero within the uncertainty of the measurements, but of course it is not exactly zero. Before computing $\overline{\Theta}'(\varphi)$ from (9), we apply a small, globally uniform correction to $\overline{N}_\infty'$, such that after the correction the global mean of $\overline{N}_\infty'$ is exactly zero. As can be seen from (9), this is sufficient to ensure that $\overline{\Theta}'(\varphi)$ is zero at the North Pole. The correction to $\overline{N}_\infty'$ can be interpreted as compensating for the fact that the time rate of change term of (3) is not completely negligible.
because the global mean of $\overline{N_{\infty}'}$ is small but not zero. The correction also compensates for the inaccuracies of the data used to compute $\overline{N_{\infty}'}$.

We can say that the “job” of the global circulations of the atmosphere and oceans is to carry out the meridional energy transport shown in Fig. 2.4. 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 locally balanced the absorbed solar radiation. Such a hypothetical state is referred to as “radiative-convective equilibrium;” modeling studies of radiative-convective equilibrium are discussed in Chapter 6. Radiative-convective equilibrium would presumably entail much warmer temperatures in the tropics, and much colder temperatures at the poles. The global circulation of the atmosphere and oceans has a moderating effect on the global distribution of temperature, tending to warm the higher latitudes and cool the tropics. As discussed in Chapter 7, these same thermal contrasts represent a source of energy (called “available potential energy”) that makes the global circulations of the atmosphere and oceans possible.

If we consider that the global circulation of the atmosphere exists in order to produce the energy transports shown in Fig. 2.4, then we can imagine that the “strength of the circulation,” as measured for example by the total kinetic energy of the atmosphere, is determined by the magnitude of the required energy transports.

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

**Surface boundary conditions**

The global atmospheric circulation is strongly affected by the properties of the Earth’s surface, and their geographical variations. The most important properties of the Earth’s surface are as follows:

**Temperature**

The temperature of the Earth’s surface varies strongly and rapidly over land, and considerably less over the oceans. The reason for this difference between land and sea will be discussed below, in the subsection on the surface heat capacity.

The oceans cover about two thirds of the Earth’s surface. Their average depth is about 4 km. Water is heavy stuff; the mass of 1 m$^3$ of water is $10^3$ kg. The mass of the oceans is about $1.3 \times 10^{21}$ kg. For comparison, the mass of the atmosphere is about 250 times less, roughly $5 \times 10^{18}$ kg.

Not only is water dense, it has a very high specific heat: about 4200 J kg$^{-1}$ K$^{-1}$. In contrast, the specific heat of air (at constant pressure) is a little less than a quarter of that, i.e., 1000 J kg$^{-1}$ K$^{-1}$. The total heat capacity of the oceans is thus about 1000 times larger ($250 \times 4$) than the
total heat capacity of the atmosphere. When the oceans say “Jump,” the atmosphere says “How high?”

The density of the atmosphere decreases exponentially upward with height, and can change by 10% or so in the course of a year, at a given location, due to changes in temperature and pressure. In contrast, the density of sea water varies by only a few percent throughout the entire ocean; it is a complex but fairly weak function of temperature, salinity, and pressure. Because of the near-incompressibility of water, pressure effects (called “thermobaric” effects) are relatively unimportant; variations of the density are mainly due to changes in temperature and salinity. Warmer and fresher water is less dense and tends to float on top; colder and saltier water is more dense and tends to sink. Surface cooling and evaporation create dense water; surface heating and precipitation create light water. Note that the properties of the water are altered mainly near the surface; below the top hundred meters or so, the properties of water parcels remain nearly invariant, even over decades or centuries.

In the study of the atmosphere, we often treat the sea-surface temperature (SST) as a seasonally varying lower boundary condition. Fig. 2.5 shows the observed distributions of the SST for January and July. 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 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 global atmospheric 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 distribution of the SST is a major complication hindering our understanding of the atmosphere and ocean as a coupled system.

An Introduction to the Global Circulation of the Atmosphere
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 vapor pressure immediately above a wet surface, called the “saturation vapor pressure,” is a strong function of temperature only. It is approximately given by

\[ P = \frac{E}{g} \]

Figure 2.5: a) SST distribution for January. The contour interval is 2°C, with shading for values higher than 28°C. b) Same for July. c) SST difference between March and September, with light shading for values below -5°C and dark shading for values above 5°C. In each panel, zonal means are shown on the right.

**Wetness**

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 vapor pressure immediately above a wet surface, called the “saturation vapor pressure,” is a strong function of temperature only. It is approximately given by

*An Introduction to the Global Circulation of the Atmosphere*
\[ e_{\text{sat}}(T) = 6.11 \exp \left[ \frac{L}{R_v} \left( \frac{1}{273} - \frac{1}{T} \right) \right] \text{ hPa}, \]

where the temperature is given in K, \( L = 2.52 \times 10^6 \text{ J K}^{-1} \) is the latent heat of water vapor, and \( R_v = 461 \text{ J K}^{-1} \text{ kg}^{-1} \) is the gas constant for water vapor. The enormous value of the latent heat of water vapor is one of the reasons why moisture strongly influences the Earth’s climate and the global circulation of the atmosphere. The strong temperature-dependence of \( e_{\text{sat}}(T) \) is shown in Fig. 2.6. Held and Soden (2006) pointed out that for typical surface temperatures, the rate of increase is a spectacular 7% per Kelvin. Fig. 2.7 shows the geographical variation of \( e_{\text{sat}}(T) \) based on the SSTs plotted in Fig. 2.5. As discussed later, near the surface over the oceans the actual vapor pressure of the air, i.e., the partial pressure of water vapor, “tries” to be \( e_{\text{sat}}(T) \), but usually falls short by 20% or so. At any rate, the “effective wetness” of the ocean increases with the SST. The largest values of \( e_{\text{sat}}(T) \) occur in the tropics, of course, and are close to 40 hPa. In those very humid regions, about 4% of the near-surface air is water vapor.

![Figure 2.6: The saturation vapor pressure, \( e_{\text{sat}} \), as a function of temperature.](image)

The availability of land-surface moisture to the atmosphere is much more complicated, and is discussed separately below.
You will not be surprised to hear that mountains have a strong effect on the global atmospheric circulation. Fig. 2.8 shows the locations of the Earth's mountain ranges. Mountains block the wind; this can be called a *mechanical forcing*. The air can flow around a mountainous obstacle, or it can flow over. Which actually happens depends in part on the scale of motion. The distribution of surface pressure across a mountain range can exert a net force on the solid Earth, and an equal and opposite net force on the atmosphere. Chapter 5 explains how this works.

Mountains can also exert a *thermal forcing* on the atmosphere, because the surface of a mountain can have a temperature quite different from that of the surrounding air at the same height. For example, during the northern summer the Tibetan plateau produces a “warm spot” in

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*Figure 2.7: The geographical pattern of $e_{\text{surf}}(T)$ based on the SSTs shown in Fig. 2.5. The contour interval is 2 hPa. Values larger than 30 hPa are shaded.*

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*Topography*

An Introduction to the Global Circulation of the Atmosphere
the middle troposphere, and this represents an important aspect of the thermal forcing that produces the Indian summer monsoon. Further discussion is given in Chapter 8.

Finally, mountains strongly influence the geographical distribution of precipitation. Rain and snow are enhanced where topography forces the air to flow uphill, and also where surface heating on sunlit slopes promotes moist convection. On the other hand, precipitation is often reduced on the downstream side of mountainous regions, because the air tends to be sinking there, and also because the moisture content of the air can be depleted by upstream precipitation maxima.

![Figure 2.8: The Earth's orography, averaged onto a mesh with a grid spacing of 1 degree of longitude by 1 degree of latitude. The contour interval is 300 m. Values higher than 1500 m are shaded.](image)

**Heat Capacity**

The heat capacity of the Earth’s surface is the amount of energy needed to change the “skin temperature” of the surface by a given amount. Here the skin temperature, \( T_s \), is defined as the temperature of an equivalent black body that would emit infrared radiation at a rate equal to the actual infrared emission by the Earth’s surface. The heat capacity is highly variable in space, and also varies somewhat less dramatically with time. The concept of heat capacity sounds simple, but it is actually somewhat subtle. The time change of the skin temperature satisfies an equation very similar to (2), i.e.,

\[
C \frac{\partial T_s}{\partial t} = N_s - \nabla \cdot G_s.
\]

(13)
Here $C$ is the heat capacity of the surface, $N_s$ is the net downward energy flux at the Earth’s surface (due to radiation and other processes discussed later), and $-\nabla \cdot G_s$ is the horizontal energy transport “inside” the Earth’s surface. For the land, we can assume that $G_s \equiv 0$, but for the oceans we expect that energy transport by currents will lead to $G_s \neq 0$. The heat capacity, $C$, depends on the composition of the material both at and below the surface, because energy flowing into the surface can be stored through a finite depth. It also depends on the speed with which energy is transported down into the material below the surface.

In general, the oceans have very high heat capacity. The geographical distribution of SST does fluctuate seasonally, however, and varies considerably with both longitude and latitude, as shown in Fig. 2.5.

The land-surface has a much smaller heat capacity. This implies that the net surface energy flux averages to nearly zero over land, even for a single day. To understand why, note that with $C \to 0$ and $G_s \equiv 0$ (appropriate for land), Eq. (13) implies that $N_s \equiv 0$. For the ocean, with large values of $C$, daily mean values of $N_s$ can be much larger. The large heat capacity of the ocean means that it is relatively hard to change the SST, because

$$\frac{\partial T_s}{\partial t} = \frac{N_s - \nabla \cdot G_s}{C},$$

(14)
i.e., a large value of $C$ in the denominator on the right-hand side of (14) reduces $\frac{\partial T_s}{\partial t}$ for a given value of $N_s - \nabla \cdot G_s$. Because of this, the SST can for some purposes be considered as a “fixed” lower boundary condition on the atmosphere.

Albedo

The degree to which the surface reflects solar radiation obviously affects its response to the sun. The surface albedo depends on surface composition and sun angle, among other things. The ocean has an albedo close to 0.06 when the sun is high in the sky, i.e., it is quite dark. At low sun angles, however, the ocean can reflect considerably more of the incident solar radiation. The albedo of the land surface varies widely, due to differing compositions of the soil or rock at the surface, differing types and amounts of vegetation cover (discussed further below), and of course the presence or absence of snow.

Roughness

“Rough” surfaces exert a drag on the wind more readily than smooth ones. The 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.

**Vegetation**

The vegetation on the land-surface regulates the flow of moisture from the soil, as discussed further below. It also affects both the roughness and albedo of the surface. The pattern of vegetation on the land surface affects the atmosphere in very complicated ways. It is virtually impossible to depict the global distribution of vegetation types without using color, which is not possible in this book, but there are many good maps available online. Obviously, the type, density, and even the health of 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.

**Sea ice**

The distribution of sea ice (Fig. 2.9) also acts as a thermal lower boundary 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 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. Until recently, the Arctic ocean has been ice-covered all year, while the North Atlantic and the Southern Oceans have long experienced seasonal melting. Of course, the thickness of the ice also varies both geographically and seasonally, and the thickness strongly determines the insulating power of the ice. In addition, several percent of open water typically occurs, especially when the ice is thin. This open water often takes the form of cracks called “leads.” The water in the leads can be much warmer than the ice nearby, especially in winter. Under such conditions, the large-scale average sensible and latent heat fluxes can be 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.
The land-sea distribution and the locations of “permanent” (or, more accurately, non-seasonal) land ice (e.g., the ice sheets that cover Antarctica and Greenland) strongly affect the surface albedo. Over land, 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 is mainly confined to Antarctica and Greenland, in the present climate, although there are many smaller glaciers throughout the world. The Greenland and Antarctic ice sheets are thousands of meters thick in places, and so increase the effective topographic height of the Earth’s surface. The distribution of land ice can vary dramatically on time scales of thousands of years and longer (e.g., Imbrie and Imbrie, 1979; Crowley and North, 1991).

Energy and moisture budgets of the surface and atmosphere

Some aspects of the global atmospheric circulation can be regarded as more or less direct responses to the various boundary conditions mentioned above. Examples include the equator-to-pole energy flux by the atmosphere, 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, which include winter storms, tropical cyclones, and many other things.

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, as shown in Table 2.2. The numbers in this table are known to two significant digits.

Of the 239 W m$^{-2}$ that is absorbed by the Earth-atmosphere system, 161 W m$^{-2}$ is absorbed by the Earth’s surface. Thus only about 239 - 161 = 78 W m$^{-2}$ of solar radiation is absorbed by the atmosphere. That is only about 1/3 of the total solar radiation absorbed by the Earth-atmosphere system.

<table>
<thead>
<tr>
<th>Absorbed solar (SW)</th>
<th>161 W m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Downward infrared (LW↓)</td>
<td>333 W m$^{-2}$</td>
</tr>
<tr>
<td>Upward infrared (LW↑)</td>
<td>-396 W m$^{-2}$</td>
</tr>
<tr>
<td>Net longwave (LW)</td>
<td>-63 W m$^{-2}$</td>
</tr>
<tr>
<td>Net radiation (SW + LW)</td>
<td>98 W m$^{-2}$</td>
</tr>
<tr>
<td>Latent heat (LH)</td>
<td>-80 W m$^{-2}$</td>
</tr>
<tr>
<td>Sensible heat (SH)</td>
<td>-17 W m$^{-2}$</td>
</tr>
</tbody>
</table>

Table 2.2: Components of the globally and annually averaged surface energy budget, after Trenberth et al. (2009). A positive sign means that the surface is warmed.

The surface receives a total (LW↓ + SW; see the notation defined in Table 2.2) of 494 W m$^{-2}$ of “incoming radiation. Note that LW↓ is about twice as large as SW! This is the “greenhouse” effect on the surface energy budget.
The incoming energy due to longwave and solar radiation absorbed by the surface is given back to the atmosphere in the form of LW↑, LH and SH. By far the largest of these is LW↑. The oceans can transport energy from one place to another, so 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 laterally at any significant rate, and their limited heat capacity forces near energy balance, everywhere, on time scales of a few days at most.

Table 2.2 shows that the net radiative heating of the surface, which amounts to 98 W m⁻², is balanced primarily by evaporative cooling of the surface at the rate of 80 W m⁻². In other words, the surface cools itself off by evaporating water.

The globally averaged energy budget of the atmosphere is shown in Table 2.3. An 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 adjusts to allow the necessary infrared emission.

*The net radiative cooling of the atmosphere, at the rate of -98 W m⁻², is primarily balanced by the latent energy source due to surface evaporation.* Of course, the 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 has to be 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 globally averaged rate of precipitation, and the globally averaged rate of evaporation, are measures of the “speed” or intensity of the hydrologic cycle. The preceding discussion suggests a second interpretation of the atmospheric energy budget: To a first approximation, the *speed of the hydrologic cycle 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

<table>
<thead>
<tr>
<th>Absorbed solar radiation (240 - 161)</th>
<th>78 W m⁻²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Net infrared cooling (-239 + 63)</td>
<td>-176 W m⁻²</td>
</tr>
<tr>
<td>Net radiative heating</td>
<td>-98 W m⁻²</td>
</tr>
<tr>
<td>Latent heat input</td>
<td>80 W m⁻²</td>
</tr>
<tr>
<td>Sensible heat input</td>
<td>17 W m⁻²</td>
</tr>
</tbody>
</table>

Table 2.3: The globally and annually averaged energy budget of the atmosphere, obtained by combining the numbers in Tables 2.1 and 2.2. A positive sign means that the atmosphere is warmed.
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, because precipitation systems produce high, cold clouds (see below) that reduce the infrared emission to space.

The local rate of precipitation is controlled mainly by dynamical processes, and the rate of evaporation from the Earth’s surface is influenced by the surface wind speed. To some extent, the overall strength of the global circulation of the atmosphere is determined by, or at least must be consistent with, the speed of the hydrologic cycle that is required to balance the globally averaged rate of atmospheric radiative cooling.

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.

In combination, these points suggest a negative feedback loop which tends to regulate the strength of the hydrologic cycle. To see how this works, consider 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. The radiative cooling acts to promote cloud formation through moist convection, so when the radiative cooling rate decreases, cloud activity slows down. In this way, the initial perturbation is damped. Further discussion is given in Chapter 5.

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, which are discussed in later chapters. In short, the atmospheric circulation carries energy upward as well as poleward.

An Introduction to the Global Circulation of the Atmosphere
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.

An example of the seasonal variations of the surface shortwave and longwave radiation at a particular station is given in Fig. 2.10, which shows the variations of the upward and downward shortwave (SW) and longwave (LW) near-surface radiation at a field site in Oklahoma. The data cover the three years 2006 - 2008. The seasonal cycle is clearly evident. High frequency
fluctuations are primarily due to cloudiness. Note that the upward solar radiation has occasional maxima during the winter. These are associated with the increased albedo of the ground following snow storms.

Data like those shown in Fig. 2.10 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 (e.g., Wielicki et al., 1996). Panel a) of Fig. 2.11 shows the meridional distribution of the zonally averaged solar radiation absorbed by the Earth’s surface, as a function of latitude, and for January, July, and the annual mean. Seasonal changes are clearly visible and easy to interpret. Near 50°N in July there is a slight dip or “shoulder” in the meridional profile of the surface absorbed solar radiation. This is associated with cloudiness, and indicates that the clouds are having a major impact on the energy budget of the ocean in those latitudes. Cloudiness also leads to a weak tropical minimum, just north of the Equator. The annual mean curve is fairly symmetrical about the Equator, but shows a minimum near 10°N associated with tropical rain systems. 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 panel b) of Fig. 2.11. In January, the strongest cooling occurs over Antarctica, and in the subtropics of the winter hemisphere. The weakest cooling occurs in cloudy regions, e.g. over the Southern Oceans and in the storm tracks of both hemispheres. Although the surface temperature is warmer in summer than in winter, at some latitudes the net longwave cooling of the surface is actually stronger in winter than in summer! 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 sometimes overwhelms the increase in the upward component, giving a net decrease in surface infrared cooling from winter to summer.

Panel c) of Fig. 2.11 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. It follows that the surface must cool by non-radiative means.

Panel d) of Fig. 2.11 shows the zonally averaged net latent heat flux. Positive values represent a moistening of the atmosphere and a cooling of the surface. 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.
Panel e) of Fig. 2.11 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.
Segue

This chapter provides a brief overview 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 energy transports in more detail, and as well as the meridional transports of angular momentum and moisture.

Among the most important points in this chapter 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 also mentioned the important effects of water vapor and clouds on the Earth’s radiation budget. Finally, we discussed the lower boundary conditions on the global atmospheric circulation associated with the distributions of continents and oceans, sea surface temperature, topography, vegetation, and ice and snow.

With this preparation, we are now ready to take a look at some of the observed features of the global circulation.
Problems

1. 
   a) Prove that, for any vector \( \mathbf{Q} \),
   \[
   \int_S \nabla \cdot \mathbf{Q} \, dS = 0,
   \]
   where the integral is taken over a closed surface, e.g., the surface of a sphere. We assume that \( \mathbf{Q} \) is everywhere tangent to the surface, i.e., it “lies in” the surface and so can be described as a “horizontal” vector. Eq. (13) shows that the globally averaged divergence of any horizontal vector is zero.

   b) Also prove that
   \[
   \int_S k \cdot (\nabla \times \mathbf{Q}) \, dS = 0,
   \]
   where the integral is taken over a closed surface. Here \( k \) is a unit vector everywhere perpendicular to the surface. Eq. (14) implies that the global mean of the vertical component of the vorticity is zero.

2. 
   a) Suppose that 1 W m\(^{-2}\) is supplied to a column of water 100 m deep. Assume that the temperature changes uniformly with depth throughout the entire column. 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 at the top of the atmosphere were used to warm the ocean uniformly, how long would it take for the temperature of the entire ocean to increase by 1 K?