Chapter 1

Energy Flows in the Climate System

1.1 Simple model of global radiation balance

The earth receives energy from the sun in the form of visible, near-infrared, and ultraviolet radiation. Most of this energy is either reflected back to space by clouds and other bright surfaces (about 30%), or is absorbed by the earth's surface. A significant fraction of the near-infrared component is absorbed by water vapor. However, since the greatest amount of vapor is found at low levels near the surface, that is where the near-infrared is preferentially absorbed in the atmosphere. Ultraviolet radiation, which forms only a small fraction of the energy coming from the sun, is mostly absorbed by ozone in the stratosphere.

Figure 1.1 illustrates a simple model of the gross radiation balance of the earth. The solar flux at the earth's orbit is $F_s = 1370$ W m^{-2}. If a fraction $A = 0.3$ is reflected by the earth, then the amount of solar radiation absorbed by the earth is equal to $F_s(1 - A)$ times the projected area of the earth πR^2, where $R = 6370$ km is the earth's radius. This amounts to 122 PW. (One petawatt = 1 PW = 10^{15} W.)

This input of energy is balanced by the outflow of thermal radiation. The radiative temperature of the earth T_{rad} is the temperature which would result in this outflow if the earth radiated like a black body:

$$\pi R^2 F_s(1 - A) = 4\pi R^2 \sigma T_{rad}^4 \tag{1.1}$$

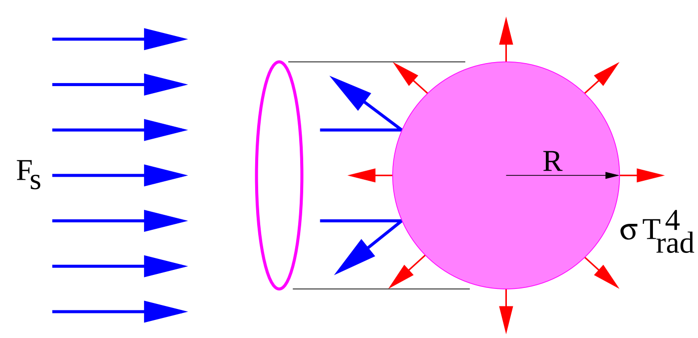

Figure 1.1: Gross radiation balance of the earth. Incoming solar radiation with flux F_s must be balanced by losses from reflection and infrared black body emission.

where $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4}. Solving for T_{rad} results in

$$T_{rad} = \left[\frac{F_s(1-A)}{4\sigma}\right]^{1/4}, \tag{1.2}$$

which for the given numbers yields $T_{rad} = 255$ K.

The radiative temperature thus calculated is significantly colder than the mean temperature of the earth's surface, which is near 288 K. This is because the earth's atmosphere is somewhat opaque to the thermal infrared radiation emitted by the surface of the earth, which means that the actual radiation to space occurs not from the surface, but from some higher level in the atmosphere, which is colder. The opacity of the atmosphere to thermal infrared radiation has little to do with the main atmospheric constituents, nitrogen and oxygen, but is caused primarily by the trace gases carbon dioxide, water vapor, methane, and especially in the stratosphere, ozone. In the troposphere, water vapor is the most important of these gases. The elevation of the surface temperature above the radiative temperature is called the *greenhouse effect*.

1.2 Stability of earth's climate

Equilibrium does not necessarily imply stability. Just because the earth is near a state of balance between incoming and outgoing energy doesn't mean that it will remain near this state. We now explore two simple "toy" models of energy flow which exhibit instability. The idea here is not to assert that the earth behaves in the manner indicated by the model, but only to indicate the breadth of possibilities.

1.2.1 Runaway greenhouse effect

As noted above, the difference between the surface temperature of the earth and its radiative temperature is due to the blocking of the radiative loss to space from the earth's surface by greenhouse gases in the atmosphere. Water vapor plays a large role in this blocking effect. However, as the surface temperature increases, the amount of water vapor in the atmosphere is likely to increase. It is reasonable to assume that the relative humidity of the atmosphere would remain unchanged as the surface gets warmer. In this case the column-integrated water vapor per unit area would be proportional to the saturation vapor pressure of water vapor at the surface temperature. Under these conditions, the greenhouse effect of water vapor would increase with the surface temperature.

The ocean covers a large part of the earth's surface and it stores most of the thermal energy in the climate system. An approximate equation for the time rate of change of mean ocean temperature T_{ocean} (assuming that the ocean covers the entire globe) is

$$\rho_w D C_l \frac{dT_{ocean}}{dt} \equiv F_n = \frac{F_s(1-A)}{4} - \sigma(T_{ocean} - \delta T_G)^4 \tag{1.3}$$

where F_n is the net flux imbalance, ρ_w is the density of sea water, D is the mean depth of the ocean (or the ocean layer being considered), C_l is the specific heat of sea water,

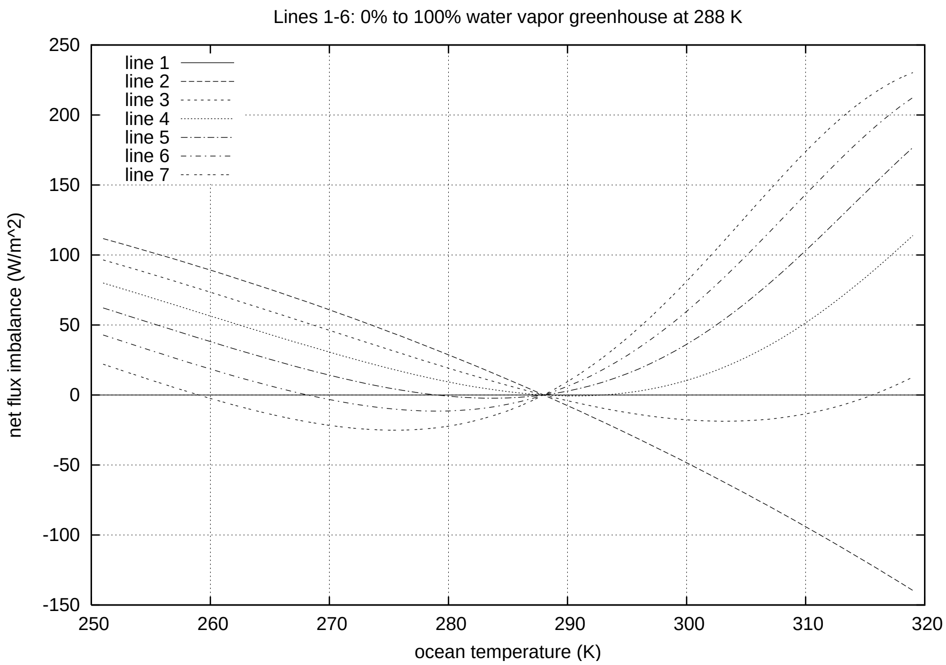

Figure 1.2: Net flux imbalance as a function of ocean temperature for fixed albedo $A = 0.3$. Lines $2-7$ represent values of the water vapor fraction $f = 0.0$, 0.2, 0.4, 0.6, 0.8, and 1.0.

and $\delta T_G = T_{ocean} - T_{rad}$ is the greenhouse gain, i. e., the difference between the ocean temperature and the earth's radiative temperature. In the above steady state model we found that $\delta T_G \equiv \delta T_{GR} = 33$ K.

Let us assume that f is the fraction of the greenhouse gain which is caused by water vapor at the reference ocean temperature T_R, which we take to be 288 K. This part of the greenhouse gain we set proportional to the saturation vapor pressure at the temperature of the ocean, while the remaining part we assume to be independent of temperature. An approximate formula for the saturation vapor pressure of water vapor is (see the next chapter)

$$e_s(T) \approx e_s(T_R) \exp\left[\frac{L}{R_v}\left(\frac{1}{T_R} - \frac{1}{T_{ocean}}\right)\right] \tag{1.4}$$

where T_R is a constant reference temperature, L is the latent heat of condensation for water, and $R_v = R_{univ}/m_w$, where R_{univ} is the universal gas constant and m_w is the molecular weight of water vapor. We therefore set

$$\delta T_G = (1-f)\delta T_{GR} + f\delta T_{GR} \exp\left[\frac{L}{R_v}\left(\frac{1}{T_R} - \frac{1}{T_{ocean}}\right)\right], \tag{1.5}$$

which yields $\delta T_G = \delta T_{GR}$ when $T_{ocean} = T_R$. Inserting this into equation 1.3 yields an expression for the net flux imbalance F_n.

Figure 1.2 shows a plot of the net flux imbalance as a function of ocean temperature for fixed albedo $A = 0.3$ and variable water vapor greenhouse fraction f. Intersection of a curve with the $F_n = 0$ line indicates an equilibrium point. The stability of this equilibrium is given by the slope of the line passing through this point – a negative slope indicates that a higher than equilibrium temperature results in cooling while a temperature lower than equilibrium results in heating. In this situation the ocean temperature relaxes back toward the equilibrium point, corresponding to stable equilibrium. On the other hand, a positive slope means that a temperature slightly warmer than equilibrium causes heating, etc., which corresponds to unstable equilibrium. Unstable equilibrium results in a runaway effect, either toward further heating or cooling, depending whether the initial temperature perturbation from equilibrium is positive or negative.

In this model the current equilibrium energy balance of the earth is unstable if the water vapor greenhouse fraction $f > 0.4$. In such a situation a slight increase in ocean temperature results in further increases without bound as the column-integrated water vapor increases and the atmosphere becomes more opaque to infrared radiation. This is the runaway greenhouse effect. Notice that a slight decrease in this case causes the system to evolve toward a stable equilibrium at a lower ocean temperature.

Clearly the current climate system is not in a state of unstable equilibrium, since any small perturbation to the system would have led long ago to either a runaway greenhouse or to a return to the lower temperature stable equilibrium point. According to this theory, the current climate can only be stable if the water vapor greenhouse fraction $f < 0.4$. This constitutes a bit of a mystery, as we believe that water vapor has a larger role than this in the greenhouse effect. It turns out that this model is significantly over simplified – we shall return to this issue later.

1.2.2 Ice world

So far we have assumed that the albedo of the earth is constant. However, the albedo can change as the amount of clouds and the amount of snow or ice cover changes. One hypothesis is that the ice age earth is a stable state because the albedo of an ice-covered earth should be much higher than that of the earth as it is today. A high albedo would have the effect of reducing the absorbed incoming solar radiation, resulting in a lower radiative temperature for the earth. We will now insert a simple temperature-dependent formula for albedo into equation (1.3) with a fixed greenhouse gain $\delta T_G = 33$ K:

$$A = 0.45 - 0.3 \tan^{-1}[(T_{ocean} - 270)/5]/\pi. \tag{1.6}$$

The albedo in this case asymptotes to 0.6 for $T_{ocean} \ll 270$ K, and to 0.3 for $T_{ocean} \gg 270$ K, with $A = 0.45$ at $T_{ocean} = 270$ K, which is near the freezing point of the ocean.

Figure 1.3 shows how the net flux imbalance varies with ocean temperature in this case. Equilibrium states exist at three temperatures, 259 K, 273 K, and 284 K. However, only the first and the third of these are stable equilibrium points – the second is unstable, as inspection of figure 1.3 shows. This system is therefore bistable, and with sufficiently strong pushes in the right direction, can be induced to switch from one state to the other. Our current state is near the upper equilibrium point. We call the lower equilibrium point "ice world", since the earth would be covered by glaciers under these conditions. This state may have something to do with the ice ages.

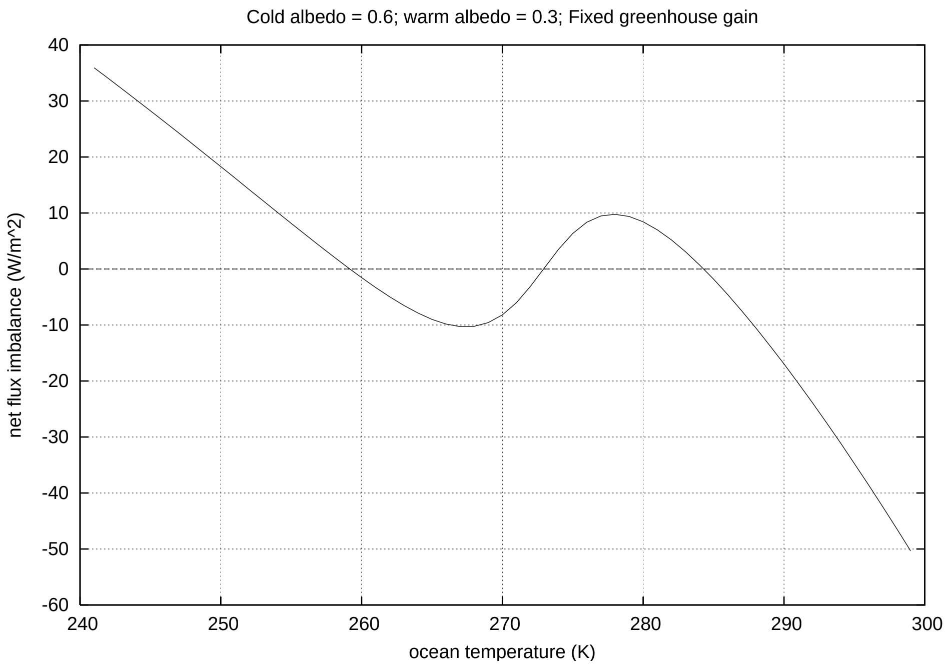

Figure 1.3: Net flux imbalance for the case of variable albedo, equal to 0.6 for $T_{ocean} \ll 270$ K and to 0.3 for $T_{ocean} \gg 270$ K.

1.3 Latitudinal energy transfer

Energy deposition is not uniform with latitude. Less solar radiation is deposited at high latitudes than low, and in the winter than in the summer hemisphere. A *local* equilibrium temperature can be computed at each latitude, but the resulting temperature distribution has a much steeper decline toward the poles than is observed. Thus, energy must be transported from the tropical regions toward the poles.

Let us make a quantitative calculation of this effect for the case in which the sun is directly over the equator, i. e., at the equinox. The key issue is the actual versus the projected area of a latitudinal strip of the earth's surface, as illustrated in figure 1.4. The actual area of the strip of earth's surface illustrated in this figure is $\delta S = 2\pi R \cos\phi \cdot R\delta\phi$, while the projected area of this strip as seen from the sun is $\delta S_p = 2R\cos\phi \cdot R\cos\phi\delta\phi$. Assuming albedo A at latitude ϕ, the energy balance at this latitude is $F_s(1-A)\delta S_p = \sigma T_{rad}^4 \delta S$, resulting in a radiative temperature there of

$$T_{rad}(\phi) = \left[\frac{F_s(1-A)\cos\phi}{\pi\sigma}\right]^{1/4}. \quad (1.7)$$

Figure 1.5 shows the latitudinal distribution of radiative temperature as well as the global radiative temperature and the mean sea surface temperature as a function of latitude. Also plotted is the sea surface temperature minus 40 K, slightly greater than the difference between the surface temperature and the radiative temperature in the globally uniform case. This can be taken as an approximation of the actual local radiative temperature. Within 50° of the equator the predicted radiative temperature exceeds the actual value, whereas

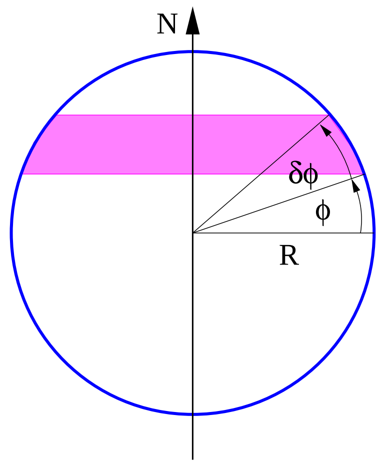

Figure 1.4: Sketch of geometry used in the calculation of the latitudinal dependence of radiative temperature at the equinox.

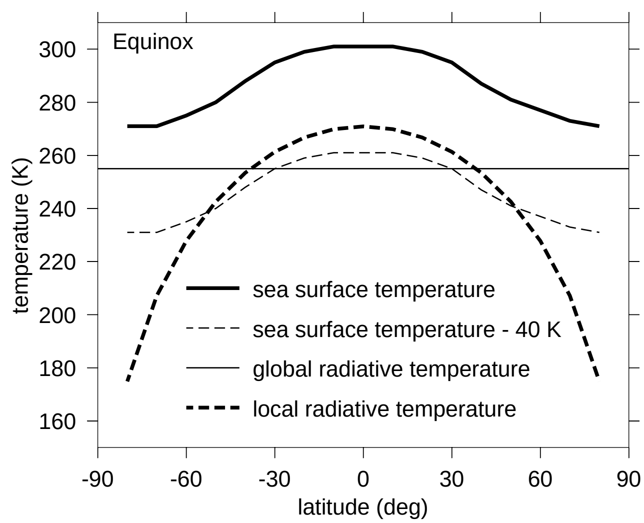

Figure 1.5: Annual and longitudinal mean of sea surface temperature, sea surface temperature -40 K, global mean radiative temperature and latitudinal distribution of radiative temperature at the equinox.

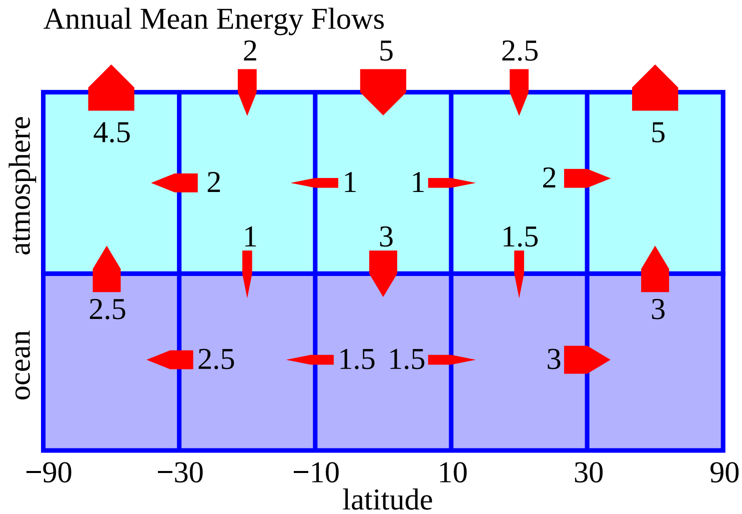

Figure 1.6: Annually averaged net energy flows in petawatts (10^{15} W; fluid transport plus radiation) between space and various latitudinal segments of the atmosphere and ocean. Data adapted from Peixoto and Oort (1992).

at higher latitudes the reverse is true. This implies lateral export of energy by the oceans and atmosphere from low latitudes to high, i. e., transport of energy down the temperature gradient. In other words, there is a net flow of energy into the atmosphere and oceans at low latitudes, followed by transport to high latitudes, where there is net export.

Figure 1.6 shows that this transport actually does take place. On an annual average, import of energy by radiation at the top of the atmosphere exceeds export by $2 + 5 + 2.5 = 9.5$ PW between 30°S and 30°N. the same amount is exported to space at higher latitudes. The poleward transport of energy is shared almost equally by the atmosphere and the ocean, with the ocean contributing slightly more.

Within 10° of the equator, the atmospheric absorption of solar radiation and the emission of infrared are nearly in balance, so that the net absorption is only about 2 PW. The absorbed energy is exported to higher latitudes. This compares with a solar input of $F_s(1 - A) \cdot 2R^2 \cos^2 \phi\delta\phi$, which equals 27 PW for $\phi = 0$ and $\delta\phi = 20/57.3$radians. Thus, the atmosphere in this band exports laterally only about 7% of the incoming solar radiation. However, an additional 11%, or 3 PW of incoming solar energy travels indirectly to higher latitudes via the oceans.

The transition between net inflow from space to net outflow to space occurs near latitudes ±30°. This is lower in latitude than suggested by the estimate in figure 1.5 where this transition occurs nearer 50°. However, we must remember that figure 1.5 is based on rather loose arguments.

Figure 1.7 shows the global energy flows as in figure 1.6, except averaged over December, January, and February only, i. e., during the northern winter. As would be expected from the southerly position of the sun during this period, there is a net inflow of energy into the southern hemisphere and a net outflow in the northern hemisphere. These hemispheric imbalances are partially compensated by flow of energy from south to north in both the atmosphere and the ocean. However, this flow doesn't account for all of the southern hemi-

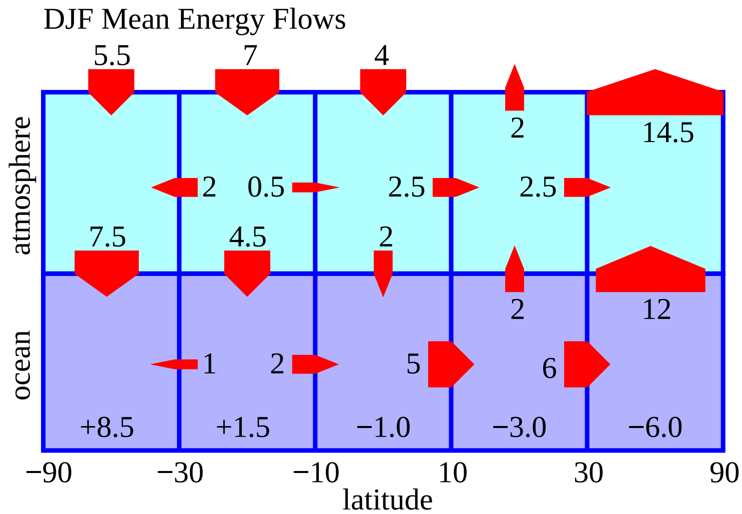

Figure 1.7: December-January-February averaged net energy flows (in petawatts; fluid transport plus radiation) between space and various latitudinal segments of the atmosphere and ocean. The numbers at the bottom are the rates at which energy is stored in the ocean segments. Zero storage is assumed for the atmosphere. Data adapted from Peixoto and Oort (1992).

sphere gains and northern hemisphere losses. Substantial warming with time occurs in the southern hemisphere as well as substantial cooling north of the equator. This heat storage effect is most important in the oceans, as the oceanic heat capacity is much higher than that of the atmosphere. The situation for the northern summer is nearly a mirror image of that for the northern winter.

1.4 Vertical energy transfer

1.4.1 Atmosphere

As noted above, solar energy is deposited at the surface and at low levels in the atmosphere, while departing thermal infrared radiation is emitted on the average at atmospheric temperatures of 255 K, which corresponds to an average elevation of about 6 km. Two processes transport this energy upward, thermal radiation and convection. It turns out that radiative transfer is incapable by itself of performing the necessary transport, because the resulting vertical temperature profile is unstable to convection below a certain level. The layer in which convection occurs in the atmosphere is called the *troposphere*. The convectively stable layer above the troposphere is called the *stratosphere*, and the boundary between the two layers is called the *tropopause*.

In the stratospheric layer the atmosphere is close to *radiative equilibrium*, i. e., the vertical radiative energy fluxes are such that zero net heating occurs at each level in this layer. We showed in the previous section that the tropical atmosphere within 10° of the equator has small lateral fluxes compared to the upward and downward energy fluxes due to solar radiation, thermal radiation, and convection. Thus, the predominant balance in the tropical troposphere is nearly one of *radiative-convective equilibrium*, i. e., one in which convective heating at each level is balanced by radiative cooling.

Since the surface of the earth is nearly covered by water, the earth's atmosphere contains a great deal of water vapor. Since rising convective parcels cool in their ascent, they eventually reach a level, called the lifting condensation level (LCL), at which the water vapor in the parcel begins to condense. This has three consequences. First, the condensation releases latent heat, resulting in higher temperatures in ascent than in a parcel without condensation. Second, much of this condensed water falls out of the parcel as precipitation. As a result, the parcel follows a different thermodynamic trajectory in the descending part of its convective cycle. Third, the small water drops and ice crystals which form when a parcel reaches the LCL and the freezing level interact strongly with solar and thermal radiation, thus modifying the radiative transfer of energy in the atmosphere. The net result is a complex and interesting system with many stubborn scientific uncertainties of importance to weather and climate.

1.4.2 Ocean

Water is highly opaque to infrared radiative transfer, and even in the clearest ocean water most of the solar radiation is deposited in the upper 100 m or so. Convection occurs in the ocean in regions where the high density of cold, salty surface water is sufficient to produce negative parcel buoyancies. This sinking water becomes the source of deep ocean water. The upward return circulation occurs at low latitudes far from the sinking motion, which occurs only in preferred regions at high latitudes. The complete circulation system is called the thermohaline circulation. Our knowledge of the details of this circulation is quite sketchy at this point. In particular, the mechanisms governing the upwelling of deep water are not well understood.

Just as the budget of moisture is inextricably intertwined with the budget of energy in the atmosphere, the budget of salt plays a crucial role in the energy budget in the ocean. This is because salt content has a large effect on ocean water density. Salt content is increased as a result of evaporation at the ocean surface, and is decreased by the inflow of fresh water into the ocean from rivers and precipitation.

The salt content of the ocean is measured in terms of *salinity* S. Salinity is defined officially in terms of the electrical conductivity of ocean water. However, in practical terms the salinity is just the mass of dissolved salts per unit mass of water, measured in grams of salt per kilogram of fresh water.

1.5 References

Marshall, J., and R. A. Plumb, 2008: *Atmosphere, Ocean, and Climate Dynamics: An Introductory Text.* Elsevier, Amsterdam, 319 pp.

Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate.* American Institute of Physics, New York, 520 pp.

1.6 Laboratory

1. Radiometer and greenhouse effect: Use the radiometer to determine the solar energy per unit area per unit time reaching the surface and the downwelling thermal radiation flux during the night.

1.7 Problems

1. Given the cooling rate of the ocean north of 30°N in figure 1.7, estimate how much the ocean cools in the three month period covered by the figure. The annual temperature cycle in the ocean is limited to roughly the upper 100 m of the ocean, thus only this layer should be considered. In doing this calculation you will have to estimate the ocean area north of 30°N.

2. Solution to equation (1.3) for the case in which $f = 0$ and δT_G is constant:

   (a) Find the steady-state value of T_{ocean}, which we shall call T_0.

   (b) Approximate the right side of equation (1.3) by assuming that T_{ocean} only differs slightly from T_0, so that the right side may be linearized in $T' = T_{ocean} - T_0$. Obtain the general solution $T'(t)$ for this equation, which is in the form of an exponential decay: $T'(t) = T'(0)\exp(-t/\tau)$.

   (c) Assuming a mean ocean depth of $D = 2000$ m, compute the time constant τ.