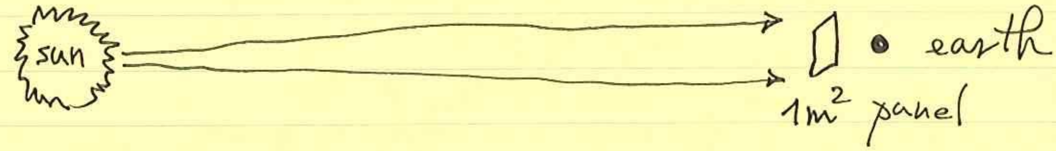
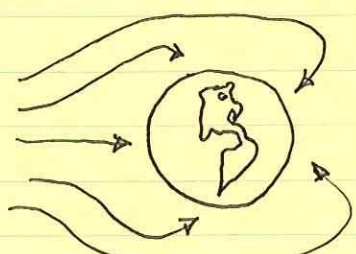
The Earth's surface temperature is governed by solar radiation:



Solar constant : $\Omega = 1360 \ W/m^2$

The average solar flux at the top of the atmosphere (including the night side) is



$\Omega/4 = 340 \ W/m^2$

The spectral content (color) of sunlight is very close to that of black body.

The energy emitted per meter^2 with wavelength between λ and $\lambda + d\lambda$ by a black body is described by the Planck distribution:

$$dU = \frac{8\pi h c \lambda^{-5}}{\exp\left(\frac{hc}{kT\lambda}\right) - 1} d\lambda$$

c = speed of light $= 3 \cdot 10^{8}$ m/sec

k = Boltzmann's constant $= 1.38 \cdot 10^{-23}$ J/°K

h = Planck's constant $= 6.63 \cdot 10^{-34}$ J sec

Depends only upon the temperature T!

Temperature <u>must</u> be measured in <u>Kelvins</u>

$$\boxed{T(^\circ K) = T(^\circ C) + 273.15^\circ}$$

The total energy $\int_{\text{all } \lambda} dU$ is given by

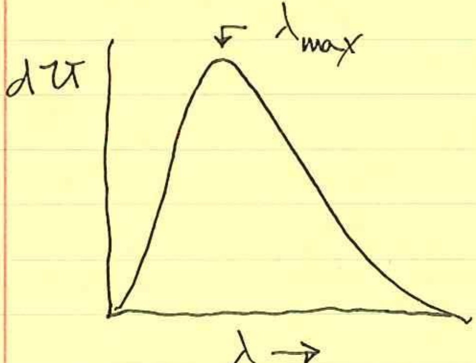
$$\boxed{U = \frac{8\pi^5 (kT)^4}{15\,(hc)^3} = \sigma T^4}$$

Stefan – Boltzmann law: energy $\sim T^4$

$$\sigma = \frac{8\pi^5 k^4}{15\,(hc)^3} = 5.67 \cdot 10^{-8} \ \frac{W}{m^2 \, {}^\circ K^4}$$

is the Stefan – Boltzmann constant

The peak λ_{max} of the Planck distribution occurs at



$$\boxed{\begin{aligned} \lambda_{max} T &= 0.2014052 \left(\frac{hc}{k}\right) \\ &= 2.90 \cdot 10^{-3} \ m \, {}^\circ K \end{aligned}}$$

Wien's law: dominant wavelength $\sim 1/T$.

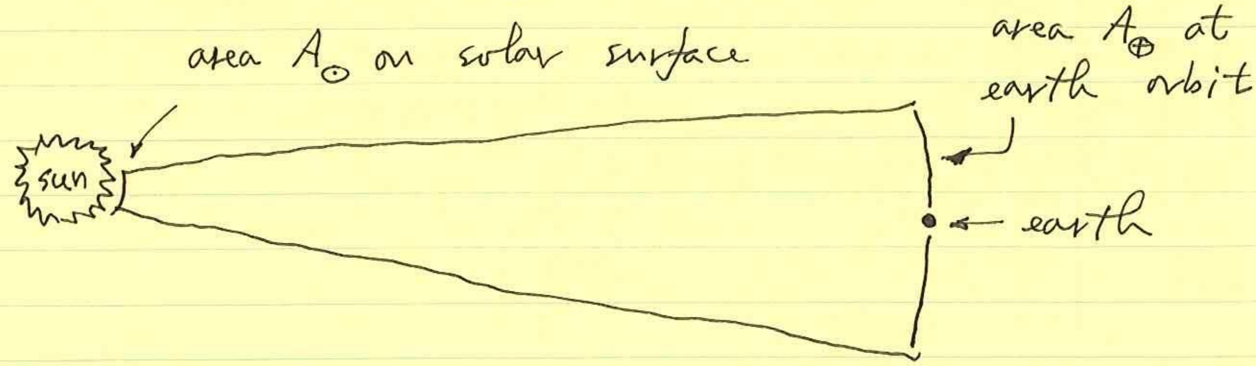
For the sun: $\lambda_{max} = 0.5\,\mu m = 5 \cdot 10^{-7}\, m$
$= 500$ nanometers

in the blue

The temperature of the solar surface is

$$T_{\odot} = 5800\ ^{\circ}K \qquad \left(\frac{2.9 \cdot 10^{-3}}{5 \cdot 10^{-7}} \right)$$

The intensity of sunlight falls off like 1/ distance^2 from the sun



$$\frac{A_{\oplus}}{A_{\odot}} = \left[\frac{\oplus - \odot \text{ distance } (1\ AU)}{\text{solar radius}} \right]^2$$

$$= \left(\frac{1.5 \cdot 10^8\ m}{7 \cdot 10^5\ m} \right)^2 = 45,900$$

Solar flux at solar surface:

$$1360 \times 45,900 = 6.2 \cdot 10^7\ W/m^2$$

Temperature found in this way is the same:

$$T_{\odot} = \sqrt[4]{\frac{6.2 \cdot 10^7}{5.67 \cdot 10^{-8}}} = 5800\ ^{\circ}K$$

Of the incoming solar radiation:

$$\boxed{100 \text{ units} = 340 \text{ W/m}^2}$$

6 units : backscattered into space by air
20 units : reflected by clouds
4 units : reflected by surface

This 30% is the short-wave (visible) radiation of ⊕ into space — what one "sees" in astronaut photos.

An additional 16 + 3 = 19 units are absorbed by water vapor, dust, clouds.

Only 100 − (6 + 20 + 4 + 16 + 3) = 51 units are <u>absorbed</u> by the ground.

The 30% that is reflected or scattered (with <u>no</u> change in wavelength) constitutes the ⊕'s <u>albedo</u>

$$\boxed{\oplus \text{ albedo}: \quad a = 0.3}$$

The remaining 70% is absorbed by the ⊕ (either the atmosphere or solid ⊕) and then re-radiated as long-wavelength infrared radiation.

Radiative balance requires that:

$$\boxed{\text{total flux in} = \text{total flux out}}$$

We can use this to calculate the temperature of a radiantly warmed $\oplus$:

flux in : $\Omega/4$
flux out : $a\Omega/4 + \sigma T_\oplus^4$

radiant temperature of $\oplus$

$$\Omega/4 = a\Omega/4 + \sigma T_\oplus^4$$

$$\boxed{T_\oplus = \sqrt[4]{\frac{(1-a)\Omega}{4\sigma}} = 255\ ^\circ K = -18^\circ C}$$

brrr!

The actual mean surface temperature of the Earth is

$$\boxed{T_\oplus = 288\ ^\circ K = 15^\circ C \quad \text{— } 33^\circ C \text{ warmer}}$$

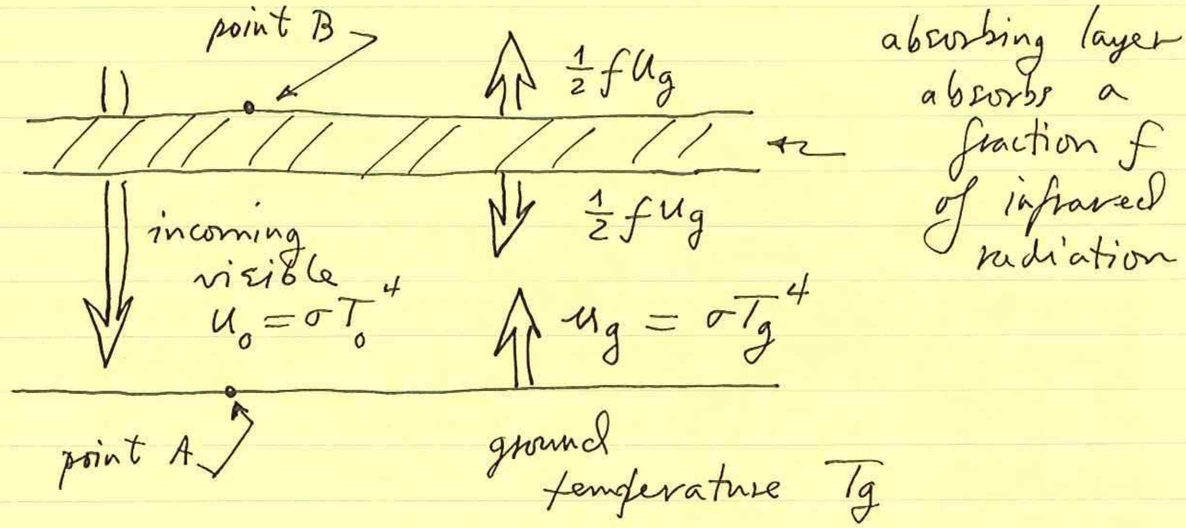
The "extra" 33°C is due to the <u>atmospheric greenhouse effect</u>.

The characteristic wavelength of the $\oplus$'s long-wave (infrared) radiation may be found from Wien's law:

$$\boxed{\lambda_{MAX} = 10\ \mu m = 10^{-5}\ m}$$

A highly simplified greenhouse warming model:

A single-layer absorber



T_0 = temperature in absence of greenhouse effect
= 255 °K

T_g = actual ground temperature

$u_0 = \sigma T_0^4$: incoming visible radiation

$u_g = \sigma T_g^4$: upward infrared radiation from ground

The absorbing layer absorbs an amount $f u_g$ and re-radiates it half upward and half downward. The downward (trapped) infrared $\frac{1}{2} f u_g$ is the source of the warming

We may examine the radiation balance condition either at point A (ground) or point B (top of atmosphere).

Point A :

$$\star \quad \underbrace{u_0 \text{ (visible)} + \tfrac{1}{2} f u_g \text{ (infrared)}}_{\text{incoming}} = \underbrace{u_g \text{ (infrared)}}_{\text{outgoing}}$$

Point B :

$$\star\star \quad \underbrace{u_0 \text{ (visible)}}_{\text{incoming}} = \underbrace{u_g (1-f)}_{\substack{\text{outgoing IR} \\ \text{from ground}}} + \underbrace{\tfrac{1}{2} f u_g}_{\substack{\text{outgoing IR} \\ \text{from atmosphere}}}$$

$\star$ and $\star\star$ agree (a good check)

$$u_0 = \left(1 - \tfrac{1}{2} f\right) u_g \text{ , or}$$

$$\boxed{\frac{T_g}{T_0} = \sqrt[4]{\frac{1}{1-\frac{1}{2}f}}}$$

To heat up to T_g (observed) $= 288\,°K$
requires

$f = 0.77$ 77% absorption in a single layer

More complete radiation balance model works in same way — treat each layer separately — known concentration of CO_2, H_2O, CH_4, etc. — these absorb IR and re-radiate <u>both</u> <u>upward</u> & <u>downward</u>.

In detail there are several other considerations:

(1) some of incoming visible is absorbed (and re-radiated as IR) by atmosphere

(2) not all IR is equally well absorbed (8–12 μm : atmosphere nearly transparent)

(3) some heat is transported upward by convection (of hot air) — the sensible heat flux

(4) some more is transported upward by evaporation of seawater — the latent heat flux

Details of IR budget:

6 IR units escape to space

108 units are absorbed by atmosphere; the net absorption is $108 - 93 = 15$ units

total IR radiated from ground = 114 units

93 of the 108 absorbed units are re-radiated to the ground (the analogue of $\frac{1}{2} f U_g$ in the one-layer model)

the net IR from the ground is $114 - 93 = 21$ units

the total incoming radiation absorbed by the ground consists of:
51 units visible + 93 units IR = 144 units

The outgoing energy from the ground also consists of 144 units:

114 units IR radiation + 7 units sensible heat flux + 23 units latent heat flux = 144 units

The total IR radiation to space consists of:

$$\left.\begin{array}{ll} 16+3: & \text{short-wave radiation absorbed by atmosphere} \\ 15: & \text{long-wave absorbed by atmosphere \& re-radiated upward} \\ 7+23: & \text{convectively transported upward} \end{array}\right\} \text{from atmosphere}$$

6 : from solid $\oplus$ (8-12 μm band)

70 units = 1 − albedo

The upshot is that the ground absorbs not 1 − albedo = 70 units but rather 114 units = 51 visible + 93 IR. It radiates the same amount upward as IR

As a result, the $\oplus$'s surface temperature is given by

$$\sigma T_g^4 = \underset{\text{not 0.7 as before}}{1.14} \times 340 \; W/m^2 = 388 \; W/m^2$$

$$\boxed{T_g = \sqrt[4]{\frac{1.14}{0.7}} \times 255°K = 288°K}$$

The atmospheric temperature drops by about 6.5°C / km with altitude.

Most of the IR radiated to space comes from mid-to-upper troposphere, which has a temperature ~ 255°K.

Thus, viewed from space, the ⊕ looks like a ~ 255°K black body.

<u>Evaporation</u>: the rate of evaporation is

23 units latent flux   surface area of ⊕

$$\frac{0.23 \times 340\ \mathrm{W/m^2} \times 5.1 \cdot 10^{14}\ \mathrm{m^2}}{2.46 \cdot 10^{6}\ \mathrm{J/kg}}$$

latent heat of H_2O
= heat required to evaporate 1 kg

$$= 1.6 \cdot 10^{10}\ \mathrm{kg/sec} = 5 \cdot 10^{17}\ \mathrm{kg/yr}$$

$$\boxed{\begin{aligned} &\text{evaporation rate} = 500{,}000\ \mathrm{km^3}\ \text{of } H_2O\text{/year} \\ &= 97\ \mathrm{cm/yr}\ \text{average over whole } \oplus \end{aligned}}$$

This is of course also the average world-wide <u>precipitation</u> rate.

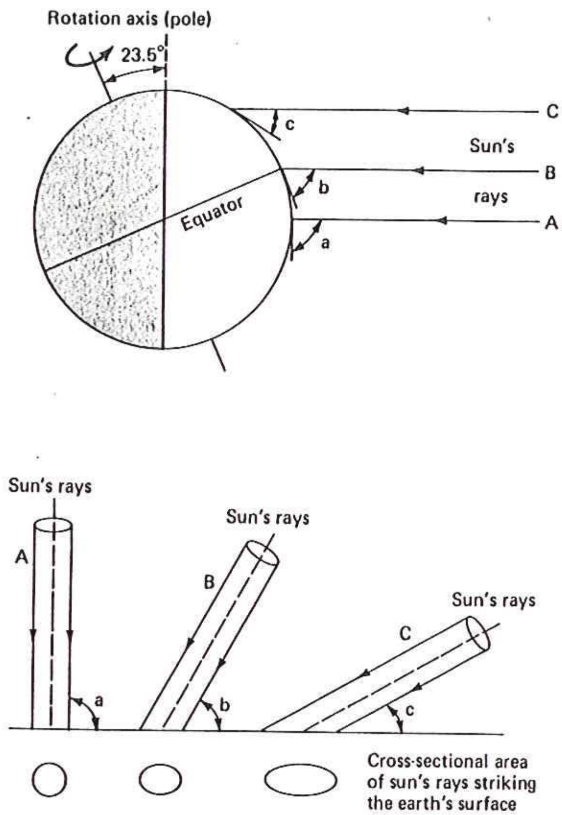


Figure 2.6 Schematic diagrams showing the variations of solar intensity (energy per unit area) with angle of incidence to the earth's surface. Lower angles (higher latitudes) result in the same energy spread out over a larger area and, thus, in a lower intensity of radiation. Scene depicted is for Northern Hemisphere winter. (Modified from A. Miller, et al. *Elements of Meteorology, 4th ed.* Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

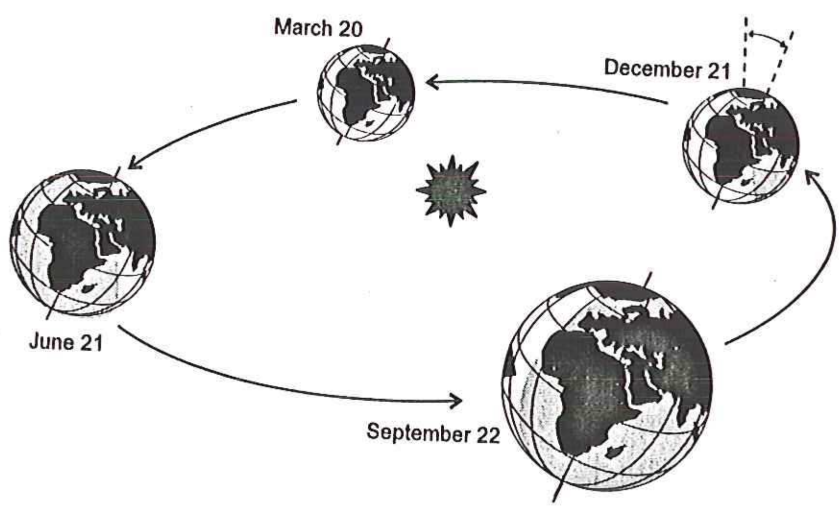


Figure 3.11. March of the seasons; as the tilted Earth revolves around the Sun, changes in the distribution of sunlight cause the succession of the seasons. (Imbrie and Imbrie 1986)

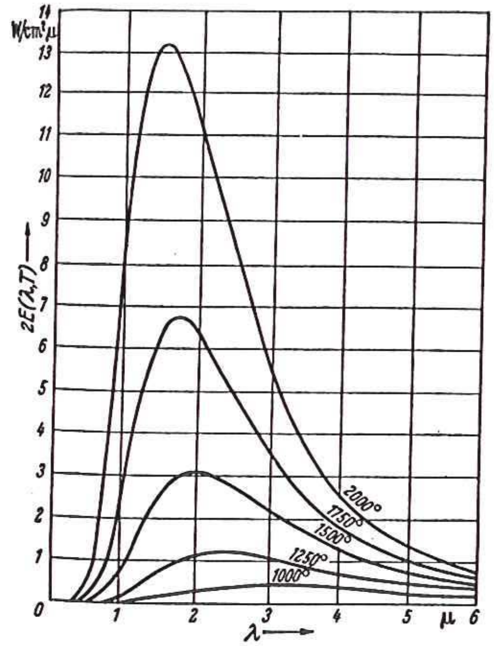
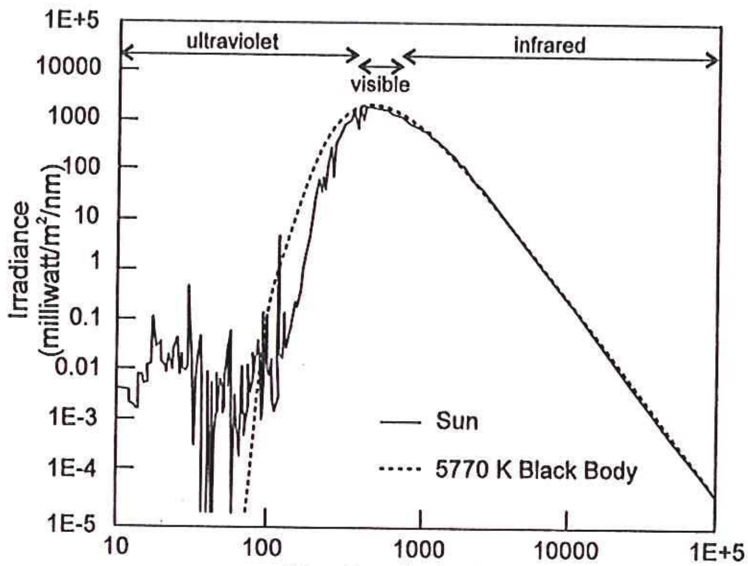
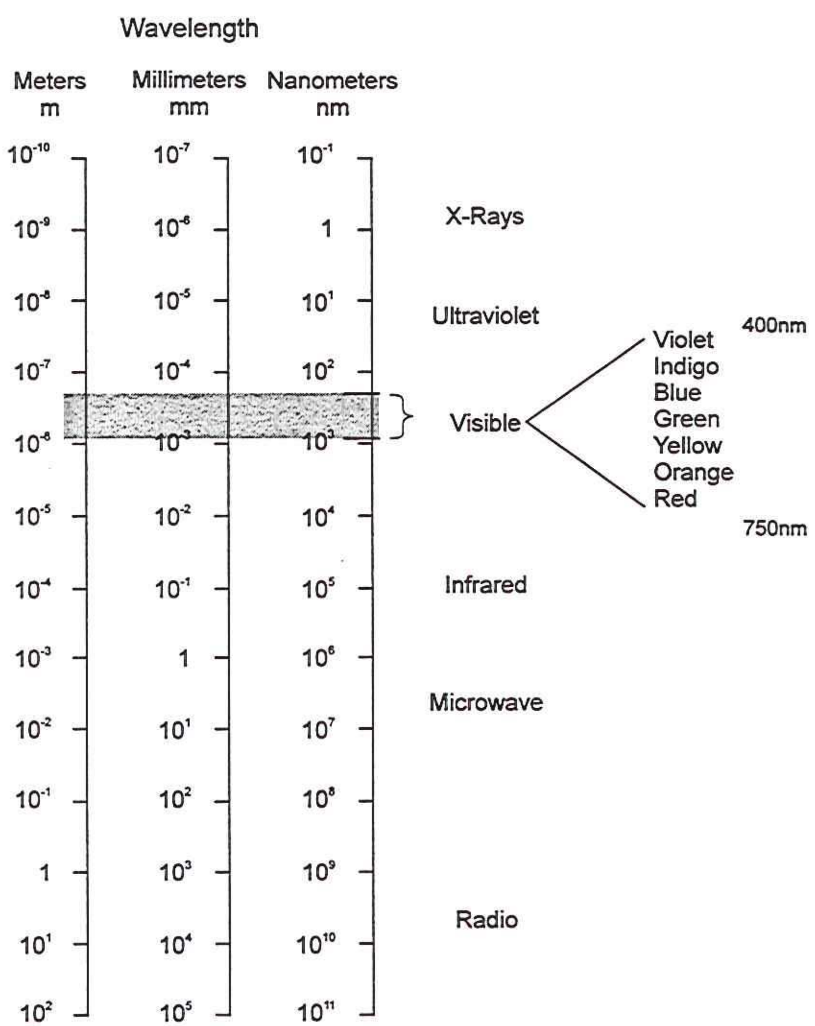


Figure 6.5. Power per unit wavelength per steradian emitted by a blackbody emitter at different temperatures. (Finkelnburg, 1964, p. 45, Fig. 20.)

Figure 2.6. The Sun's spectral irradiance typical of solar minimum conditions compared with the spectrum of a blackbody radiator at 5,770 K (Lean 1991). UV = ultraviolet; VIS = visible; IR = infrared.





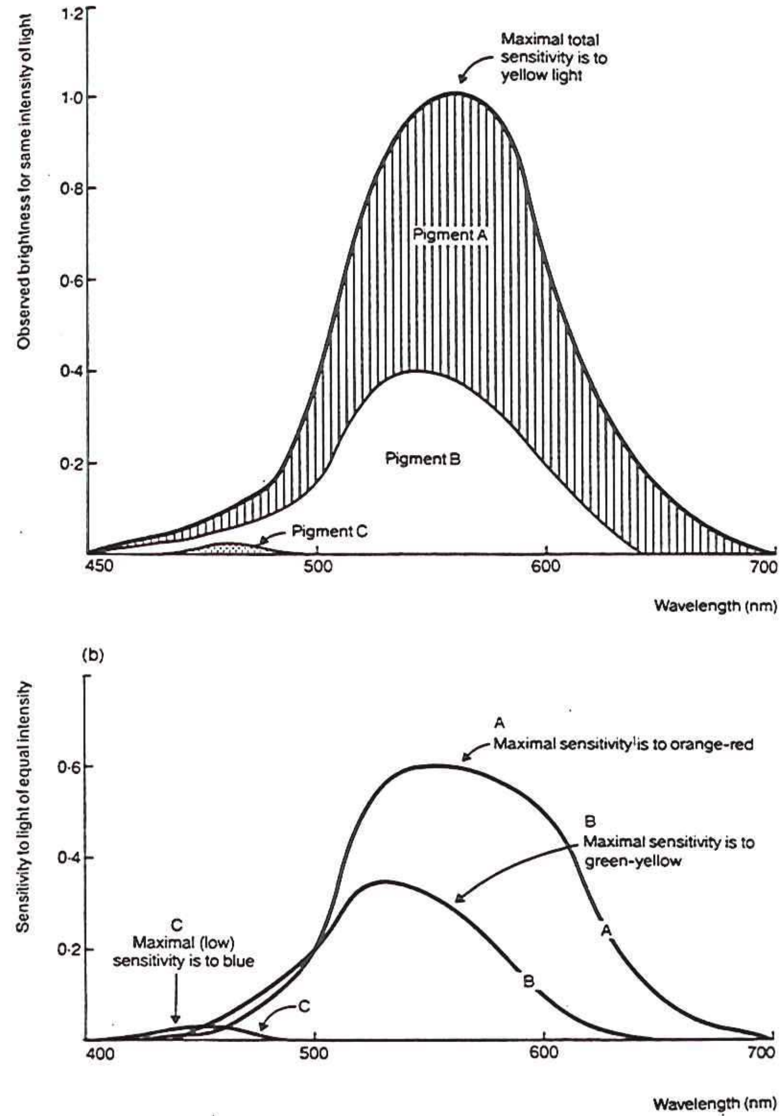


Figure 47. Brightness by day *(a) The full line shows the combined response of the cone cells to light of the same intensity but of different wavelengths. The approximate contributions of the three systems of cone pigments are indicated by the differently shaded areas. (b) The approximate sensitivities of each of the separate systems.* The valves are those estimated at the cornea by Smith and Pakorny.

Major Components (concentration in percent by volume in dry air)	
Nitrogen, N_2	78.08
Oxygen, O_2	20.95
Argon, Ar	0.93
Minor Components (concentration in parts per million by volume, ppmv)	
Water vapor, H_2O	40–40,000
Carbon dioxide, CO_2	360
Neon, Ne	18.2
Helium, He	5.24
Methane, CH_4	1.7
Krypton, Kr	1.1
Trace Components (incomplete list; concentration in parts per billion by volume, ppbv)	
Hydrogen, H_2	550
Nitrous oxide, N_2O	330
Xenon, Xe	87
Carbon monoxide, CO	60–200
Ozone, O_3	10–30
Ammonia, NH_3	4–20
Formaldehyde, CH_2O	0–10
Nitric oxide, NO	1
Nitrogen dioxide, NO_2	1
Sulfur dioxide, SO_2	1–4
Chlorofluorocarbons	
F11 ($CFCl_3$)	0.18
F12 (CF_2Cl_2)	0.38
Carbon tetrachloride, CCl_4	0.13
Methyl chloride, CH_3Cl	0.6

Sources: Holland 1978; Warneck 1988; Rowland and Isaksen 1987.

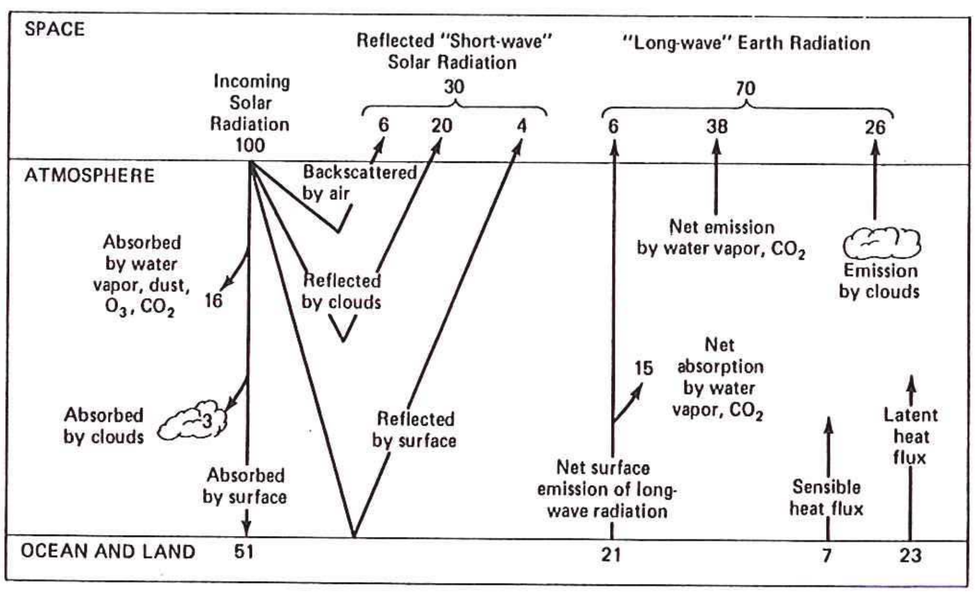


Figure 2.4 The mean annual radiation and heat balance of the atmosphere and earth. Units are assigned so that incoming solar radiation (0.5 cal/cm²/min) is set equal to 100. ("Short-wave" solar radiation is that with <4μm wave length; "long-wave" earth radiation is >4μm). (Adapted from U.S. Committee for the Global Atmospheric Research Program 1975).

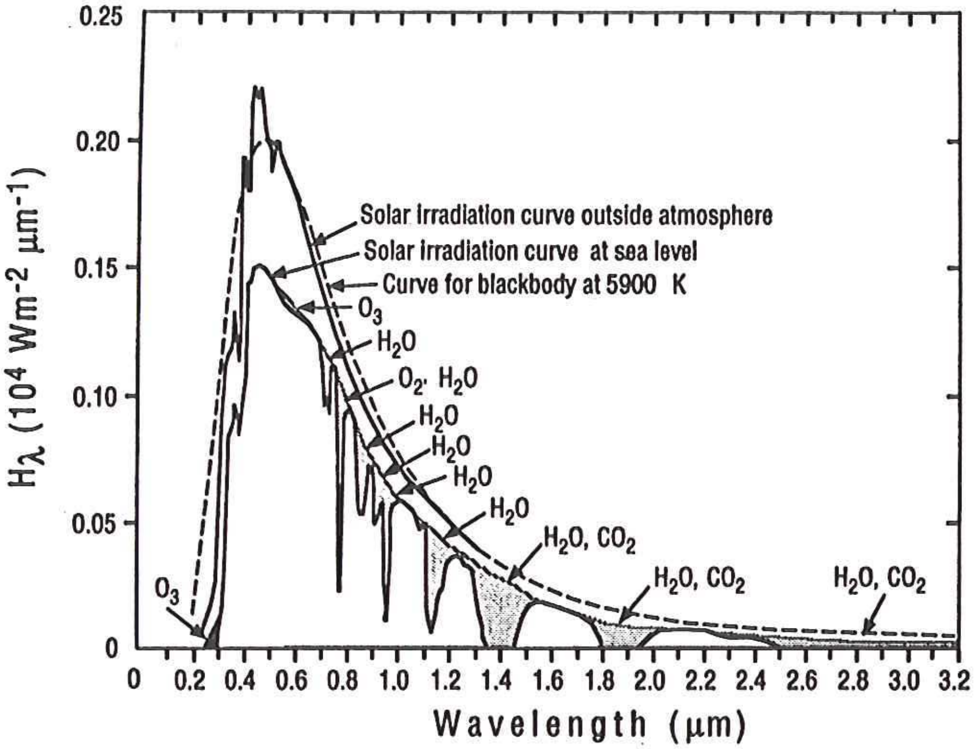


Figure 2.2 Spectral distribution of incident solar radiation outside the atmosphere and at sea level. Major absorption bands of some of the important atmospheric gases are indicated. (Reproduced by permission of McGraw-Hill from S. L. Valley (ed.), *Handbook of Geophysics and Space Environments*, McGraw-Hill, New York, 1965, Fig. 16.1, p. 16.2) The emission curve of a black body at 5900 K is shown for comparison

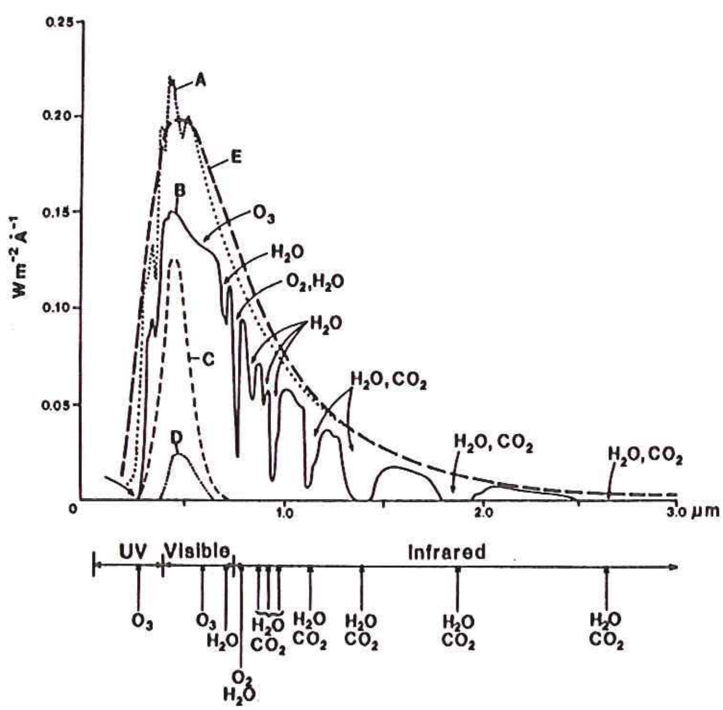


Figure 6.4. The solar spectrum (A) outside the Earth's atmosphere, (B) at sea level, (C) 10 m below the sea surface, and (D) 100 m below the sea surface; E is the spectrum of a blackbody radiating at the same temperature as the solar photosphere. The valleys in curve *B* reflect absorption by molecules in the terrestrial atmosphere. (Adapted from Dietrich et al., 1975.)

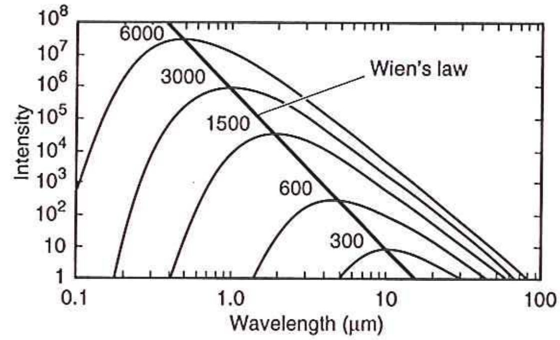
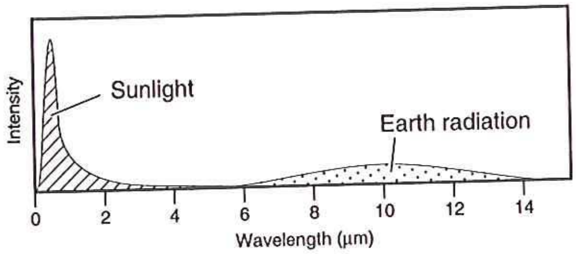


Figure 3.7 Blackbody radiation spectra as a function of temperature (kelvin), over the entire range of temperatures relevant to environmental studies. The values are displayed here on a log-log graph, so that both the wavelength and intensity scales are greatly compressed and cover many orders of magnitude. (From P. R. Gast, Air Force Cambridge Research Laboratory, McGraw Hill (1967). Appendix B of Revision of Chapter 22 of the *Handbook of Geophysics and Space Environments*, Air Force Survey in Geophysics #199, Office of Aerospace Research, USAF, Bedford, Mass. Template only.)

Figure 3.8 The relative spectra of sunlight and Earth's blackbody radiation (referred to as terrestrial radiation or Earthglow). The spectral regions of the emissions are seen to be quite distinct, with little overlap of spectra.



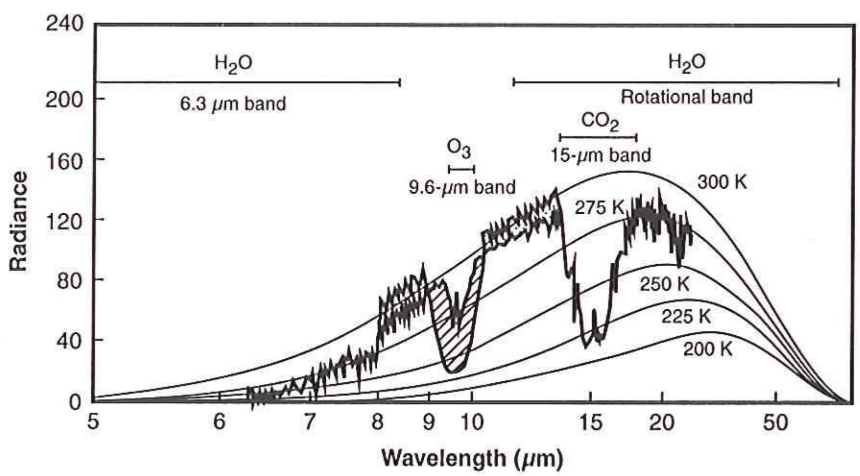


Figure 11.11 The spectrum of radiation emitted by the Earth to space, as measured from a satellite. The spectrum is the jagged line, which defines the intensity of the radiation as a function of wavelength. Several ideal blackbody-emission curves (Planck functions) corresponding to temperatures ranging from 200 K to 300 K are shown for comparison. By matching the observed emission spectrum at any wavelength to a blackbody curve, the effective temperature of the Earth's emitting region at that wavelength can be estimated. Indicated above the spectrum are the atmospheric species responsible for the emissions in the corresponding wavelength intervals. The relevant molecular "bands" for each species also are identified. The response of the emission spectrum to the addition of a hypothetical atmospheric absorber in the longwave window region is illustrated by the hatched regions. The measurement corresponds to a cloud-free area of the Earth. (Data from Hanel, R. A., B. J. Conrath, V. G. Kunde, C. Prabhakara, I. Revah, V. V. Salomonson and G. Wolford, "The Nimbus 4 Infrared Spectroscopy Experiment 1. Calibrated Thermal Emission Spectra," *Journal of Geophysical Research* 77 [1972]: 2629.)

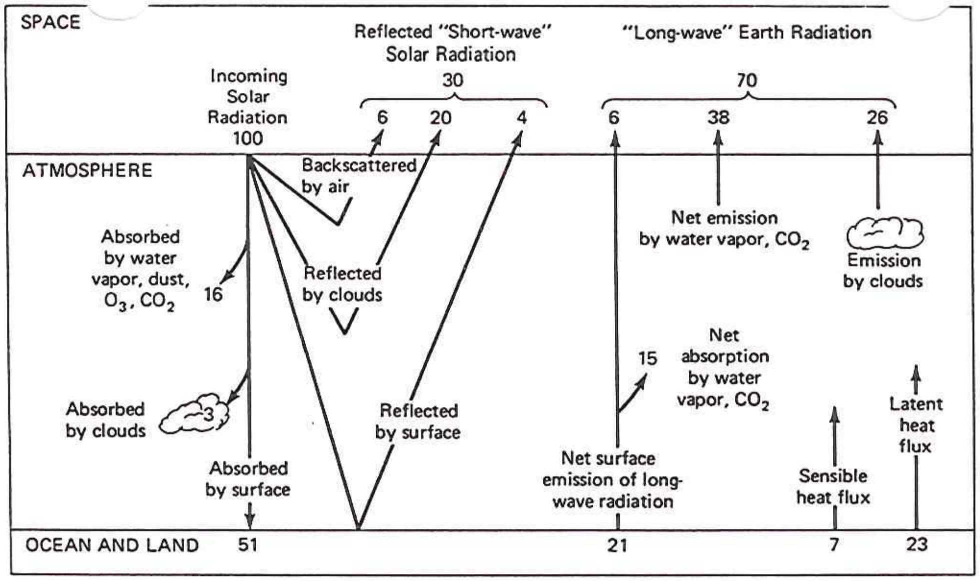
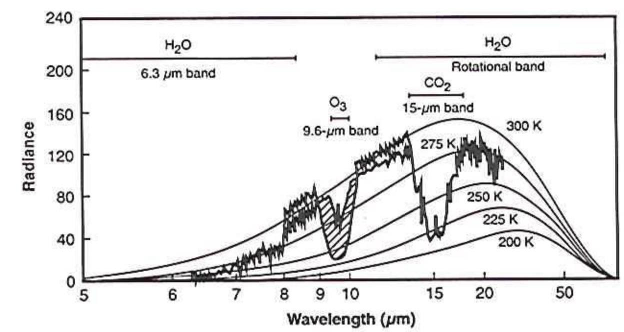


Figure 2.4 The mean annual radiation and heat balance of the atmosphere and earth. Units are assigned so that incoming solar radiation (0.5 cal/cm²/min) is set equal to 100. ("Short-wave" solar radiation is that with <4μm wave length; "long-wave" earth radiation is >4μm). (Adapted from U.S. Committee for the Global Atmospheric Research Program 1975).



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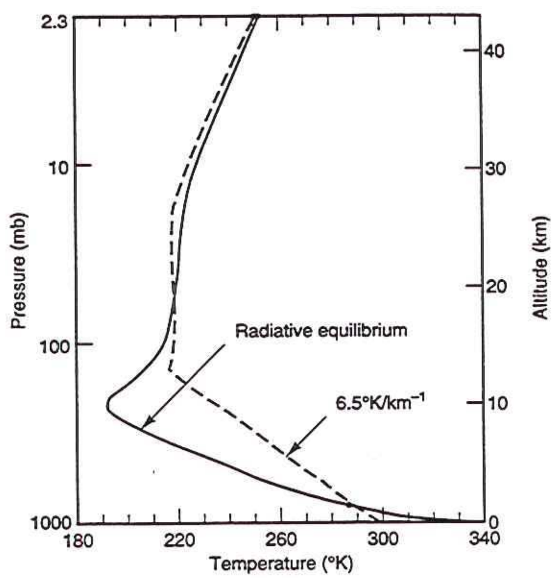
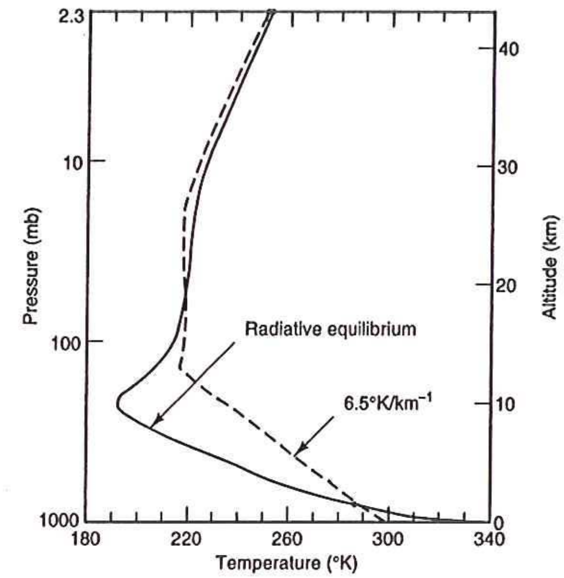


FIGURE 3–6 The variation of temperature with altitude in a radiative equilibrium model assuming: (1) pure radiative equilibrium only, and (2) convection and precipitation occurring, resulting in a lapse rate (or decrease in temperature with increasing altitude) of $6.5°K\ km^{-1}$. (After S. Manabe and R. F. Strickler, 1964, *J. Atmos. Sci.* v. 21, p. 361. American Meterological Society.)



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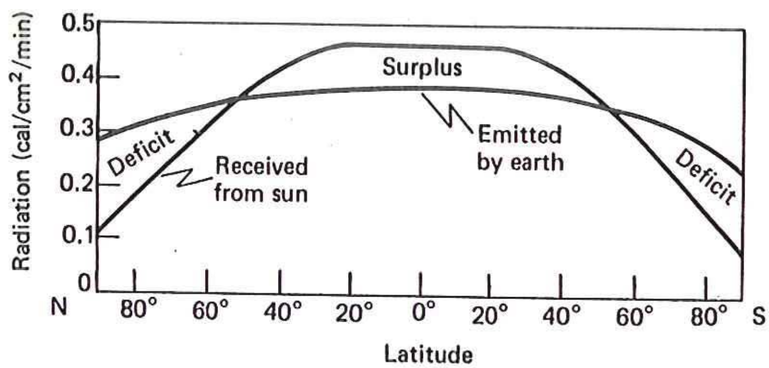

Figure 2.8 Mean annual radiation absorbed from the sun and radiated from the earth to space, as a function of latitude. (Adapted from A. Miller, et al. *Elements of Meteorology, 4th ed.* Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

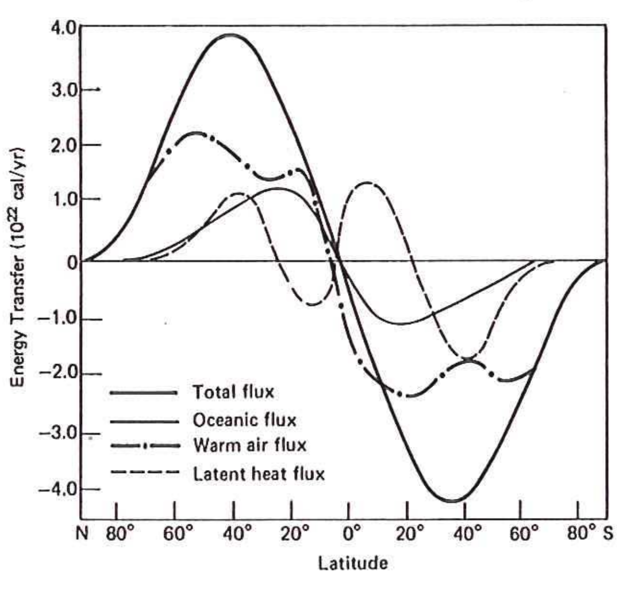

Figure 2.9 Rates of poleward energy transport by various mechanisms. Positive energy transfer values refer to northward transport and negative values to southward transport. (Adapted from W. D. Sellers, *Physical Climatology*. Copyright © 1965 by The University of Chicago Press. All rights reserved.)

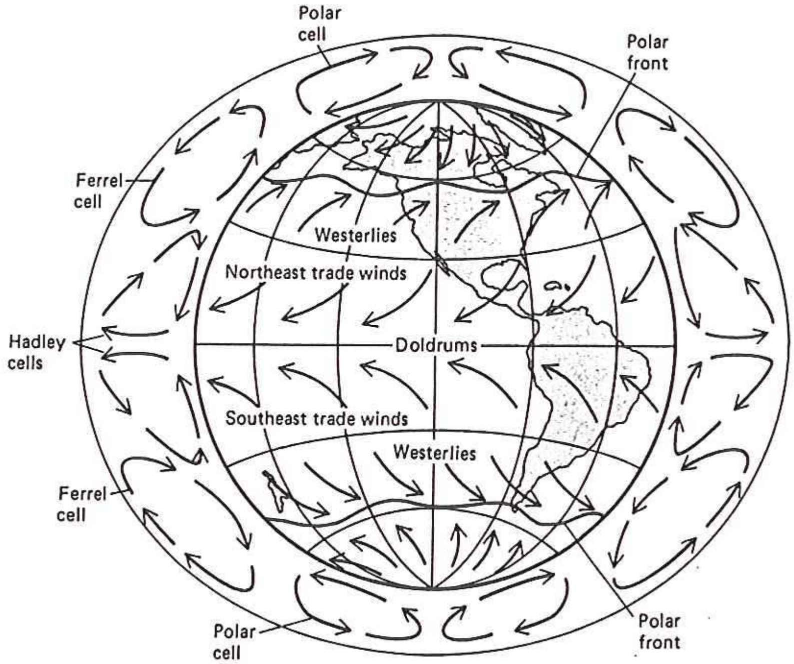

Figure 2.10 Schematic representation of the general circulation of the atmosphere. (Modified from A. Miller, et al. *Elements of Meteorology, 4th ed.* Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

TABLE 2.2 Fluxes in the Water Cycle with Methods of Determination

Process	Flux km³/yr	Flux cm/yr[a]	Source
Precipitation on land	110,300	74	Lvovitch 1973
Evaporation from land	72,900	49	(Precipitation on land minus run-off)
Runoff from land (river runoff and direct groundwater discharge to the ocean of ~ 6% of total)	37,400	25	Baumgartner and Reichel 1975; groundwater discharge, Meybeck 1986
Precipitation on oceans	385,700	107	(Total precipitation minus precipitation on land)
Evaporation from oceans	423,100	117	(Precipitation on oceans minus runoff)
Total precipitation on earth	496,000	97	Baumgartner and Reichel 1975
Total evaporation on earth	496,000	97	Baumgartner and Reichel 1975 (equal to total precipitation)

Note: Because of use of different areas, values in cm/yr do not balance between land and oceans.

[a] Fluxes in cm/yr calculated on the following basis: area of earth = 510×10^6 km² (total evaporation and precipitation); area of oceans = 362×10^6 km² (precipitation and evaporation over oceans); and area of land = 148×10^6 km² (runoff, precipitation and evaporation over land).