Temp of Earth

[redacted]

This lecture devoted to radiation: third form of energy transport: negligible in solid ⊕ but dominant in atmosphere (and space)

So much for the good news about fossil fuels.

We will [redacted] have depleted most of the easily found pumpable oil by 2100 but coal — which can be converted into liquid fuel at relatively low cost — will still be abundant

2100 is when — according to the UN "medium" projections — world population will have leveled off at about 12 billion people.

Increasing food production by a factor of 2[redacted] will be more difficult than keeping people supplied with energy in the next 100 years.

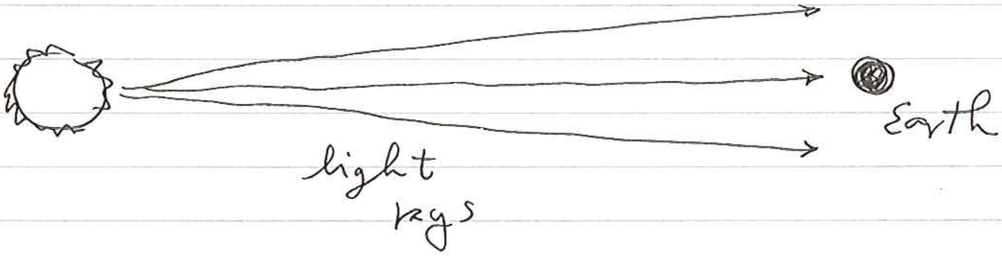
Now for the bad news — which I'm sure you know — fossil fuel burning causes global warming.

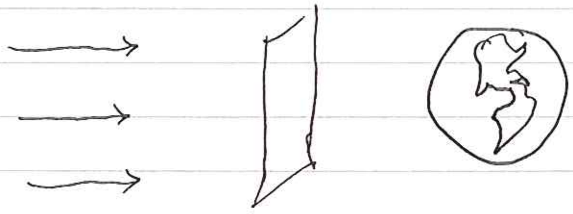
To understand this, let's back up and consider what controls the ⊕'s surface temperature.

The ⊕ is heated by the Sun.

The flux of solar energy (photons)

at the position of the Earth (solar constant) is





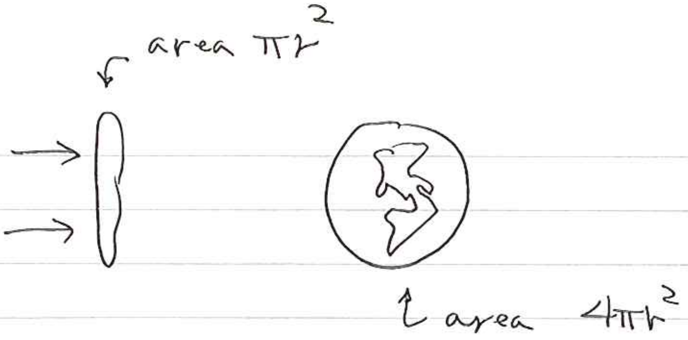
flux $\Omega = 1360$ W/m^2

The light shining through 1 m^2 in space hits 1 m^2 on ⊕ surface only if sun is directly overhead.

At other latitudes, spread out over a larger area

This is of course why it is warm in summer and cool in winter

The average solar flux onto ⊕ surface (including the night side) is



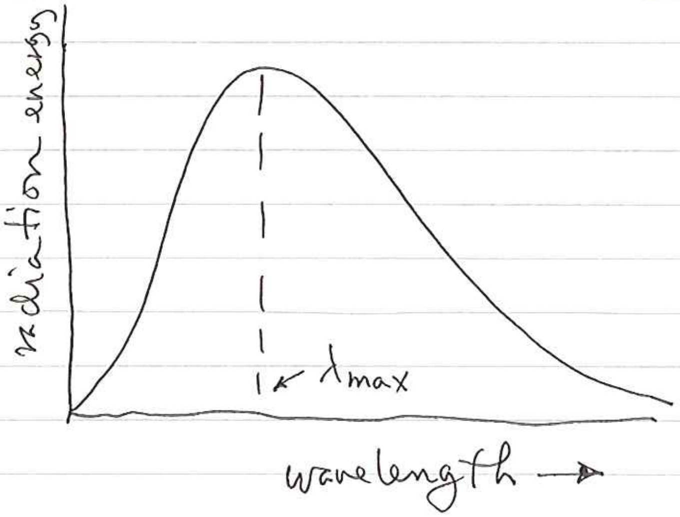
$S/4 = 340 \ W/m^2$

~~What happens to this light when it impinges on earth?~~

what does this light look like, i.e. what color is it.

The sun's radiation is very close to that of a black body, so-called because it absorbs all radiation incident upon it — a "perfect" radiator

The colors or wavelengths emitted by a black body always have the same form The so-called Planck distribution



radiant energy flux per unit wavelength

$\frac{dE}{d\lambda}$ units $\frac{W}{m^2 \cdot m}$

energy emitted by 1 m^2 of radiant surface

Wavelength of radiation emitted depends only on temperature T

$du =$ flux of energy emitted between λ and $\lambda + d\lambda$

$\frac{W/m^2}{m}$

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

Planck distribution

$c =$ speed of light $= 3 \cdot 10^5$ km/sec

$= 3 \cdot 10^8$ m/sec

$k =$ Boltzmann's constant

$= 1.38 \cdot 10^{-23}$ J / °K

$h =$ Planck's constant

$= 6.63 \cdot 10^{-34}$ J sec

The temp T is measured in Kelvin's (°K above absolute zero)

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

The total energy $\sum_{\text{all } \lambda} du$ is given by

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

Stefan — Boltzmann

$$u = \sigma T^4$$

$$\sigma = 5.67 \cdot 10^{-8} \ \frac{W}{m^2 \ {}^\circ K^4}$$

$$\sigma = 8\pi^5 k^4 / 15 (hc)^3$$

law #1 of radiation :
emitted energy (photons) $\sim T^4$

The peak of the Planck distribution occurs at

$$\lambda_{max} T = 0.2014052 \left(\frac{hc}{k} \right)$$

Wiens "displacement" law

$$\lambda_{max} T = 2.90 \cdot 10^{-3} \ m \ {}^\circ K$$

law # 2 – dominant
wavelength emitted $\sim \frac{1}{T}$

$$T_\odot = \frac{2.90 \cdot 10^{-3}}{5 \cdot 10^{-7}} = 5800 K$$

The spectrum of a hot body is sharply peaked about this maximum

The maximum for the sun is in the green

$$\lambda_{max} \cong 0.5 \ \mu m = 5 \cdot 10^{-7} \ m$$
$$= 500 \text{ nanometers}$$

The temp of the solar surface is

$$T_\odot = 5800 \ {}^\circ K$$

Problem: what is solar flux at surface of sun?

Solar radius = $7 \cdot 10^5$ km ↙ ⊕ about 200 solar radii away

$\oplus - \odot$ distance (mean) = $1.5 \cdot 10^8$ km

↑ 1 AU

$$1360 \left(\frac{1.5 \cdot 10^8}{7 \cdot 10^5}\right)^2 = 6.2 \cdot 10^7 \frac{W}{m^2}$$

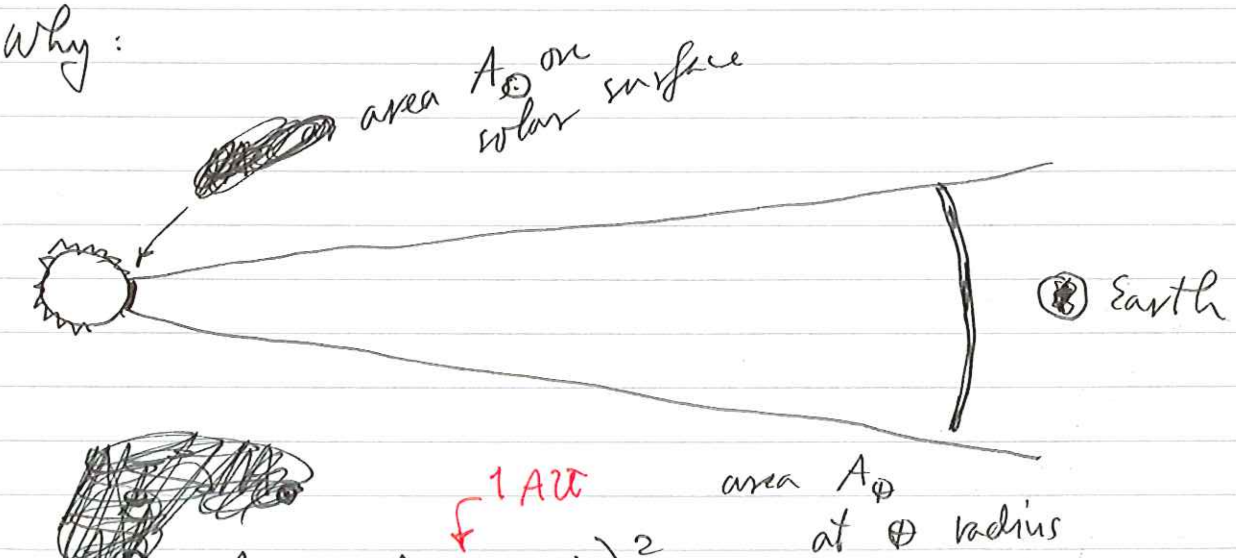
what is inferred temp of sun from Stefan — Boltzmann law

$$6.2 \cdot 10^7 = 5.67 \cdot 10^{-8} \, T^4$$

$$T = 5750 °K \text{ — good}$$

Make a good homework problem

Why:



$$\frac{A_\odot}{A_\oplus} = \left(\frac{\overset{\text{1 AU}}{\oplus - \odot \text{ dist}}}{r_\odot}\right)^2$$

intensity or energy falls off like $(1/\text{distance})^2$

53/4

Color vision — human eye has evolved so that it is sensitive to incoming solar radiation

Net absorption of pigments A, B, C in 3 types of cones ≈ solar spectrum.

Many types of color blindness but most common type due to absence of one of three cone types

(Also can be missing 2 or all 3 or have different cones)

Blue-green color blindness — missing pigment C

Chlorophyll-A : a molecule in the leaves of plants, absorbs in the red & blue.

That's why leaves are green

The most famous black body curve is from the COBE satellite (Cosmic Background Explorer)

It measured the radiant energy from deep space — remnant of the big bang.

[redacted]

$\lambda_{max} \approx 1.90$ mm ← microwave

3000 times longer than visible

$\Rightarrow T = 2.735°K$ — the temperature of the universe — has cooled since big bang

Every box is 1000σ
T universe known incredibly well!

The radiation from the sun looks very much like a black body curve with a temp $T_\odot$.

Now — what happens to this radiation as it enters the $\oplus$'s atmosphere?

Before this — ask why eyes can sense exactly the solar spectrum?

First — atmospheric composition

78 % N_2
21 % O_2
0.9 % Argon — from decay of ^{40}K

Rest are trace : 360 ppm CO_2
0.036 %

H_2O highly variable — common experience.

Of the incoming solar radiation
340 W/m^2 average

6% is backscattered into space by air

20% is reflected by clouds

4% is reflected by ⊕ surface

This is the short-wave (visible) radiation of ⊕ into space.

It's what one <u>sees</u> in astronaut photos of ⊕.

Another 16 + 3% is absorbed in the air or by clouds.

actually 51%

Only about 50% reaches the ⊕ surface

Common experience — sunbathing — sun goes behind a cloud

The absorption occurs at selective frequencies — mostly due to absorption by CO_2 and H_2O
Curve B

Curves C & D — 10 m & 100 m below sea surface — why it looks <u>blue</u>

$$\boxed{\text{albedo} = \% \text{ reflected (or scattered) with no change in wavelength}}$$

The 30% total that is reflected is the Earth's <u>albedo</u>

$$\boxed{a = 0.3}$$

viewed from outer space the ⊕ is ~ 30% as bright as the sun in the visible

The remainder 70% is absorbed either in the atmosphere or by the continents & oceans and is re-radiated as long-wavelength (infrared) radiation

Radiation balance requires that the radiant output be

$$\text{flux in} = \Omega/4 = \overset{340}{[redacted]} \ W/m^2$$

$$= \text{flux out} = a\,\Omega/4 + \sigma T_{\oplus}^4$$

↑ radiant temp of ⊕

312
290
22

$$\frac{\Omega}{4} = a\frac{\Omega}{4} + \sigma T^4$$

$$T^4 = \frac{(1-a)\,\Omega}{4\sigma}$$

$$\boxed{T = 255°K = -18°C}$$

[redacted]

this the average temp of the moon (no atmospheric greenhouse)

Actual mean surface temperature is ~ [redacted] ~15°C
= 288°K

$$\boxed{33[redacted]°C \text{ warmer}}$$

[redacted]

Possible homework problems —

What is average T of the moon

$a_{☾} = 0.07$ very dark
93% absorbed

Recall that most of ⊕'s albedo is from atmosphere
(0.26 out of 0.30)

$$T = \sqrt[4]{\frac{(1-a)\Omega}{4\sigma}} = 273\,^\circ K = 0\,^\circ C.$$

And for Mercury:

$$\Omega(\text{at Mercury}) = \Omega_\oplus \left(\frac{1}{.3871}\right)^2$$

$$\Rightarrow$$

$$T_{\text{Mercury}} = 433\,^\circ K = 160\,^\circ C$$

In fact the lit side is ~ 900°K
and the dark side is ~ 0°K.

The extra 33°C ~~[illegible]~~ warming is due to the atmospheric greenhouse effect

The wavelength of the re-radiated waves

$T = 255$ °K ⇒ $\lambda_{max} = 11$ μm

$T = 288$° K ⇒ $\lambda_{max} = 10$ μm

Infrared (long-wavelength) radiation

(Visible is 0.4 – 0.7 μm)

Infrared radiated by ⊕ into space is ~~[illegible]~~ 25 times longer λ than red = 0.4 μm.

λ_{max} for a T=288°K body is 10 μm

Can be measured using satellite IR sensors.

Looks like a 288° K radiator with "holes" in it — due to absorption of the re-radiated IR by greenhouse gases in the ⊕'s atmosphere, esp. CO_2, H_2O & CH_4.

This ~~[illegible]~~ absorbed radiation is re-radiated both up & down. Net effect is to raise temp of ⊕ surface by 35°C.

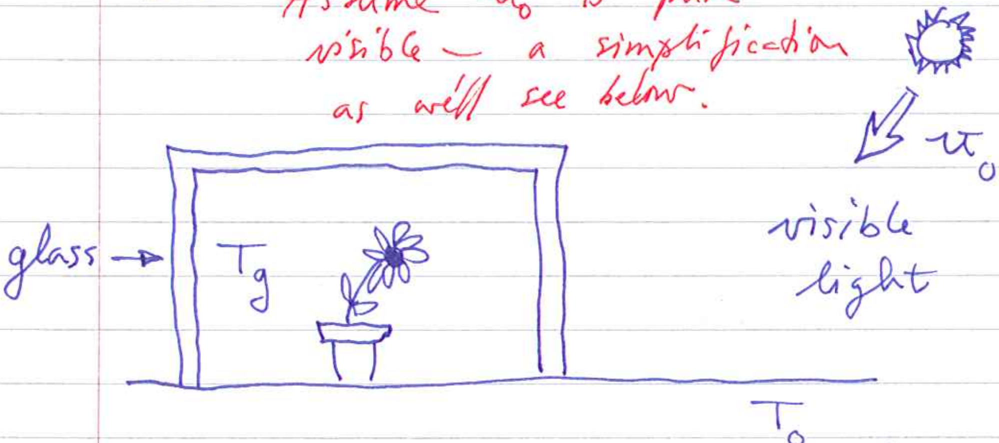
& we'd all be wearing down parkas in summer

If not for this oceans would be the world's biggest ice-skating rink.

Mechanism same as a greenhouse.

Theory simplest to understand for that case.

Assume u_0 is pure visible — a simplification as we'll see below.



depends on latitude, season, absorption by atmosphere, time of day, etc.

Outside the greenhouse — ground in radiative balance

$$u_0 = \sigma T_0^4 \leftarrow \text{IR radiated by ground}$$

The visible light passes thru the glass (it's ~~[illegible]~~ transparent — almost no absorption)

It's absorbed by plants, pots, ground inside, etc. and re-radiated as IR

The glass absorbs IR radiation.

Say it absorbs a fraction f of the radiated IR

$$u_g = \text{radiated IR} = \sigma T_g^4$$

amount absorbed / m² $= \boxed{f u_g =} \; f \sigma T_g^4$

This energy absorbed by the glass is radiated both in & out:

$\frac{1}{2} f u_g = \frac{1}{2} f \sigma T_g^4$ in each direction

Radiation balance at the ground (or plant) surface within the greenhouse requires

$$\underbrace{u_0 + \tfrac{1}{2} f u_g}_{\text{incoming}} = \underbrace{u_g}_{\text{outgoing}}$$

$$u_0 = (1 - \tfrac{1}{2} f) u_g$$

Balance on the outside of the glass requires

$$\underbrace{u_g (1 - f) + \tfrac{1}{2} f u_g}_{\text{outgoing}} = \underbrace{u_0}_{\text{incoming}}$$

$u_0 = (1 - \frac{1}{2} f) u_g$ — same: a good check

$$u_g = \frac{u_0}{1 - \frac{1}{2} f} \qquad \cancel{\sigma} T_g^4 = \frac{\cancel{\sigma} T_0^4}{1 - \frac{1}{2} f}$$

$$\boxed{T_g / T_0 = \sqrt[4]{\frac{1}{1-\frac{1}{2}f}}}$$ — greenhouse temp increase eqn

If the glass absorbs a fraction

~~f = 1/8 : T_g/T_0 = 1.034~~

~~f = 1/4 :~~

~~f | T_g/T_0~~
~~1/8 | 1.034~~
~~1/4 | 1.074~~
~~1/2~~

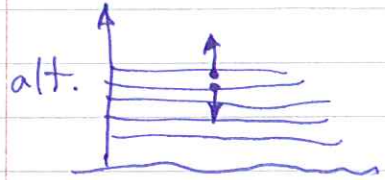
f	T_g / T_0
1/8	1.016
1/4	1.034
1/2	1.074

If $T_0 = 293°K = 20°C$:

f	$T_g - T_0$
1/8	4.7°C
1/4	10°C
1/2	22°C

} f depends upon thickness of glass

The $\oplus$'s atmosphere works in the same way — treat each level as a layer of "glass"

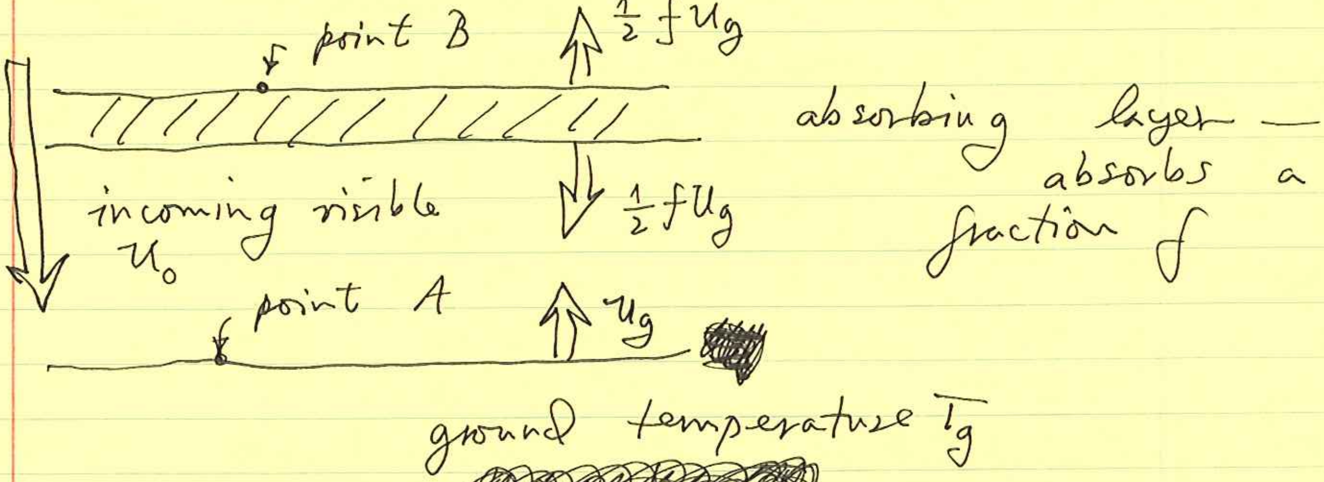


know concentration of CO_2, H_2O, CH_4 in each layer; absorb & re-radiate in both directions

A real greenhouse works primarily by another mechanism.

Traps the hot air inside — cannot rise convectively.

Because of this — describe instead as a one-layer absorber model:



incoming visible light $u_0 = \sigma T_0^4$

↑ temp in absence of absorbing layer

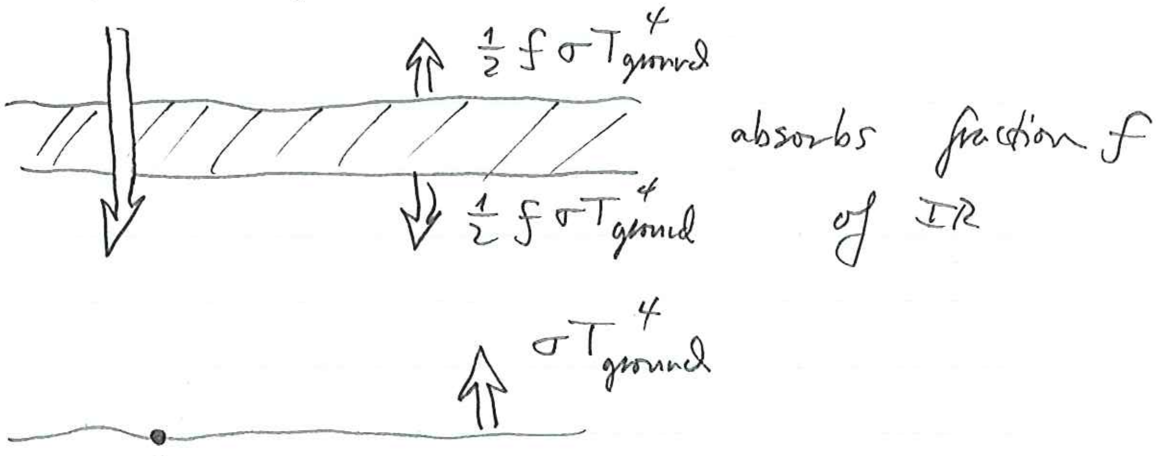
T_g = greenhouse temperature

What would f have to be to "explain" the ⊕'s temperature?

$$\left(\frac{288}{255}\right)^4 = 1.627 = \frac{1}{1-\frac{1}{2}f} \qquad \boxed{f = 77\%}$$

10 Dec 1999

$$\frac{1}{4}\Omega(1-a) = 340\,(0.7)\ \mathrm{W/m^2}$$



balance ingoing and outgoing flux here

$$\frac{1}{4}\Omega(1-a) + \frac{1}{2} f \sigma T_g^4 = \sigma T_g^4$$

$$\frac{1}{4}\Omega(1-a) = \left(1-\frac{1}{2}f\right)\sigma T_g^4$$

$$T_g = \sqrt[4]{\frac{\Omega(1-a)}{4\sigma\left(1-\frac{1}{2}f\right)}}$$

$f = 0$: $T = 255\ \mathrm{K} = -18°\mathrm{C}$

$f =$ 77% : $T = 288\,\mathrm{K} = 15°\mathrm{C}$

greenhouse warming

Total radiation received
(and re-radiated) by
ground is

call this 100 units

$$\underbrace{\boxed{\tfrac{1}{4}\Omega(1-a)}}_{\text{visible}} + \underbrace{\tfrac{1}{2} f \sigma T^4}_{\text{IR}}$$

$$= 70 + \tfrac{1}{2} f \sigma T^4$$

But $\sigma T^4 = \dfrac{70}{1-\frac{1}{2}f}$

$$70\left[1 + \frac{\frac{1}{2}f}{1-\frac{1}{2}f}\right] = \frac{70}{1-\frac{1}{2}f}$$

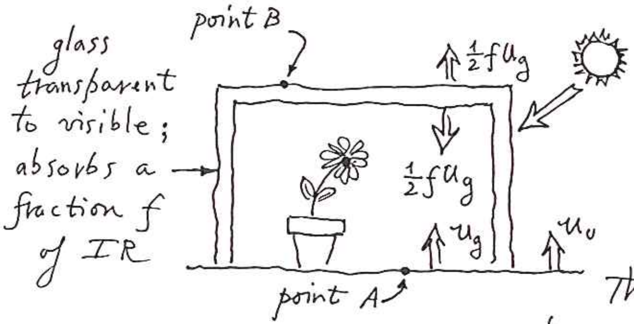
$$f = 0.77$$

Total = 114 units $= \sigma T^4_{\text{ground}}$

Same as Fig. 2.4

i.e., $\dfrac{1-a}{1-\frac{1}{2}f} = \dfrac{0.7}{1-\frac{1}{2}(0.77)} = 1.14$

Idealized Glass Greenhouse



glass transparent to visible; absorbs a fraction f of IR

incident visible radiation u_0 (W/m²) depends on time of day, season, latitude, cloud cover, etc.

The ground outside the greenhouse absorbs u_0 and re-radiates an equal amount of infrared:

$u_0 = \sigma T_0^4$ where T_0 is the ambient temperature outside of the greenhouse

The ground (and pots, etc.) inside absorbs u_0 but it radiates infrared at a higher temperature:

$$u_g = \sigma T_g^4$$

The glass walls absorb a fraction $f\sigma T_g^4$ of this and re-radiate it, half up and half down. Energy balance at point A requires:

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

Balance at point B requires, on the other hand,

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

Both equation ★ and ★★ yield the same result (a good check): $u_0 = (1 - \frac{1}{2} f) u_g$, or

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

The fraction f of IR absorbed depends upon the thickness and other properties of the glass.

f	T_g/T_0	T_g if $T_0 = 20°C = 293\,K$
1/8	1.016	4.7 °C
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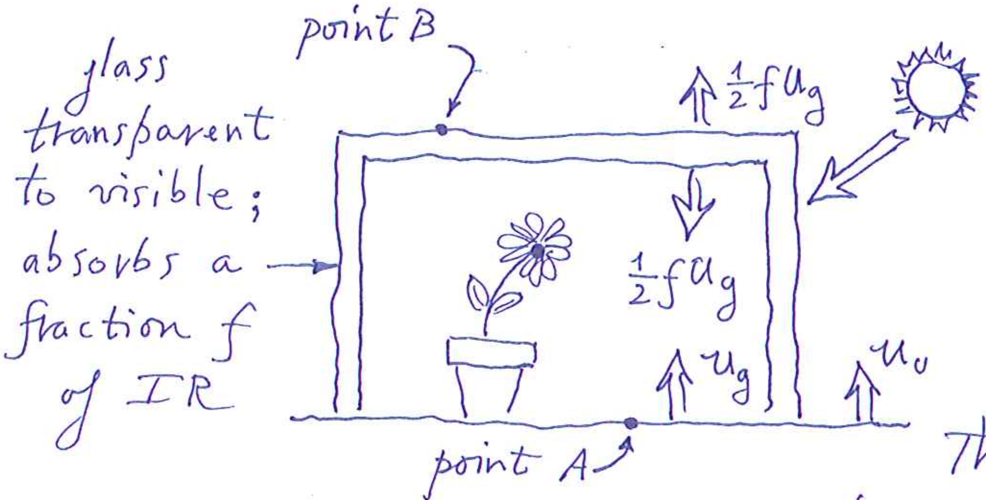

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Both equation * and ** yield the same result (a good check): $u_0 = (1-\frac{1}{2}f)\,u_g$, or

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In *detail* many other considerations:

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

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

this is the 6 units of IR out the top

(3) some heat is transported upward by convection of hot air (*sensible* heat flux) and *some* by evaporation of seawater (*latent* heat flux)

sensible — 7 of incoming units
latent — 23 of incoming units

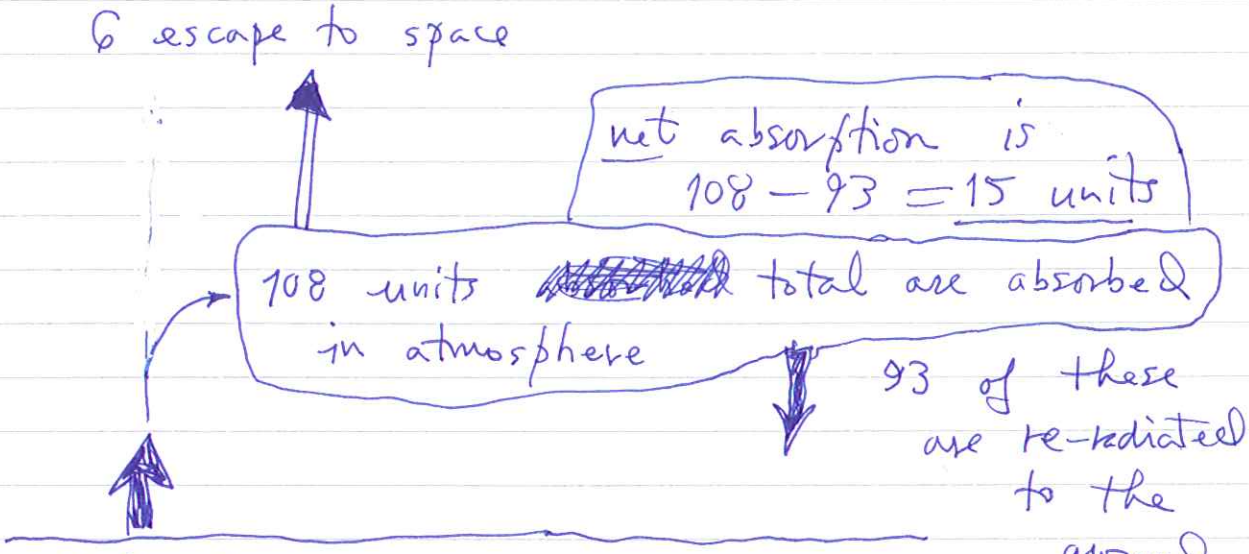
The 21 units of outgoing radiation from the ground is the *net* IR radiation

The total IR from the ground is 114 units

6 escape to space (8-12 μm band)
108 absorbed in atmosphere
↑ of this all but 15 re-radiated to ground (*93 units*)

↑ this is what causes the warming — analogous to $\frac{1}{2} f U_g$ in greenhouse case.

Details of IR budget :



93 of these are re-radiated to the ground (this is the analogue of $\frac{1}{2} f U_g$ downward in the glass greenhouse

total IR radiated from ground is 114 units

net IR from ground is 114 − 93 = 21 units

The incoming radiation absorbed by ground (surface of the ⊕) consists of 51 units visible + 93 units IR

= 144 units

Outgoing energy : 114 units IR + 7 sensible + 23 latent = 144 units

Incoming : 65% IR and 35% visible

15 units (of the 108 total) are radiated up & away to space.

The total radiated to space from the atmosphere is

$$\underbrace{16+3}_{\text{short-wave absorbed by atmosphere}} + \underbrace{15}_{\text{long-wave absorbed by atmosphere \& radiated to space}} + \underbrace{7+23}_{\text{transported to atmosphere by convection}}$$

$$= 64 \text{ units} + \underbrace{6}_{8-12\ \mu m \text{ band from solid } \oplus} = 70 \text{ total}$$

The ⊕ surface temp on this basis is

$$\sigma T_s^4 = \underset{\uparrow \text{ not } 0.7 \text{ as before}}{(1.14)} \times 340\ W/m^2 = 388\ \frac{W}{m^2}$$

$$\boxed{T_s = \sqrt[4]{\frac{1.14}{0.7}} \times 255^\circ = 288^\circ}$$

More detailed calculation "explains" observed temp structure of atmosphere

15°C at surface drops by ~6.5°C/km to −50°C at the top of the troposphere (10 km high)

Most of the IR re-radiated to space comes from the mid-to-upper troposphere, which has an average temperature of ~255°K rather than 288°K.

Thus, viewed from space, the ⊕ appears to be a ~255°K black body.

↓ latent flux

23 units of solar heat are used to evaporate H_2O from the oceans (mostly) and land

The rate of evaporation is

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

↑ latent heat of H_2O = heat required to evaporate 1 kg

evaporation rate

$$= 1.6 \cdot 10^{10}\ kg/sec = \overset{4.967}{5} \cdot 10^{17}\ kg/yr$$

$= 496{,}000\ km^3/yr$

97 cm/yr average over whole ⊕

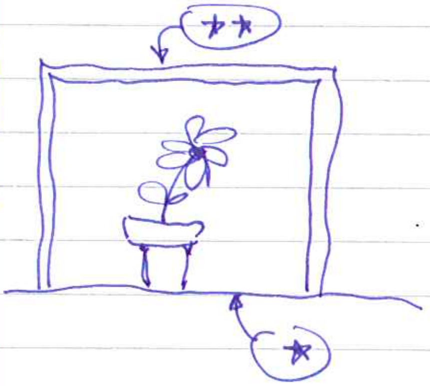
Decided not to use this calculation (for a glass greenhouse) any more.

Finally, let us re-do our greenhouse calculation, accounting for the fact that

$$i = \frac{93}{51+93} = 65\%$$ of the incoming radiation is IR

$$\boxed{u_0 = (1-i)\,u_0 + i\,u_0}$$

$$u_0 = (1-i)\cdot \underset{\text{not absorbed by glass: visible}}{u_0} + i \cdot \underset{\text{absorbed IR}}{u_0}$$



absorbed fraction:

$$A = \underbrace{f i u_0}_{\text{from outside}} + \underbrace{f u_g}_{\text{from inside}}$$

Balance at *:

$$u_g = \underbrace{u_0 (1-i)}_{\text{visible}} + \overbrace{u_0 i (1-f)}^{\text{IR}} + \underbrace{\tfrac{1}{2}(f u_g + f i u_0)}_{\text{IR radiated by glass \underline{downward}}}$$

$$\frac{u_g}{u_0} = \frac{1-\frac{1}{2}if}{1-\frac{1}{2}f}$$

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

Balance at ** :

$$u_0 = u_g\,(1-f) + \underbrace{\tfrac{1}{2}\,(f\,u_g + i f\,u_0)}_{\text{radiated upward by glass}}$$

Gives same result, as expected.

Limit $i = 0$: checks with old answer

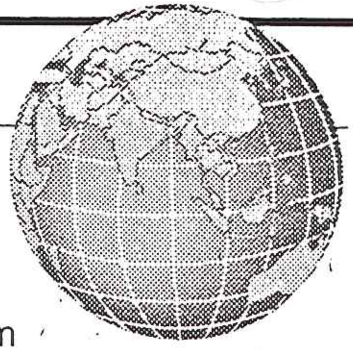
Limit $i = 1$: $T_g = T_0$ — glass no different than any other absorber

$i = 0.65$:

$$\frac{T_g}{T_0} = \sqrt[4]{\frac{1-0.32f}{1-0.5f}}$$

f	$T_0 \to T_g$
1/4	15° C → 19°C
1/2	15° C → 23°C
1	15° C → 48°C ← 118° F whew!

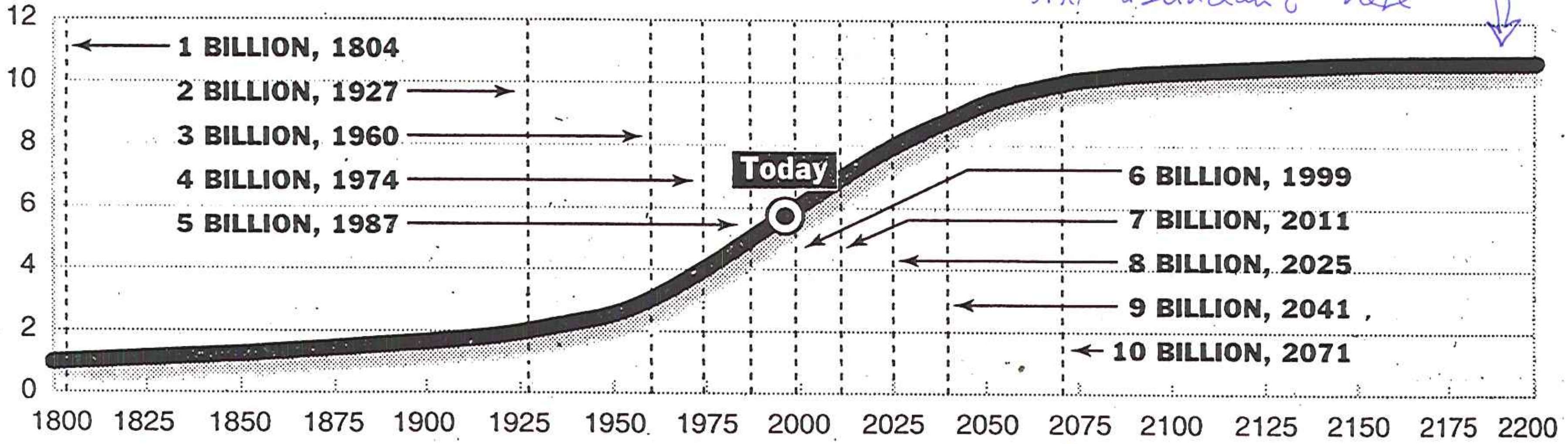
STATUS REPORT



The Population Explosion Slows Down

A new United Nations study has found that the world's population is growing more slowly than was expected. This suggests that the world's population, now 5.77 billion, will stabilize just after the year 2200 at 10.73 billion. Shown is the world population from 1800 to stabilization based on United Nations projections, in billions.

coal & oil shale still abundant here



Source: United Nations

The New York Times

plenty of fossil fuel at least until N_t approaches steady-state

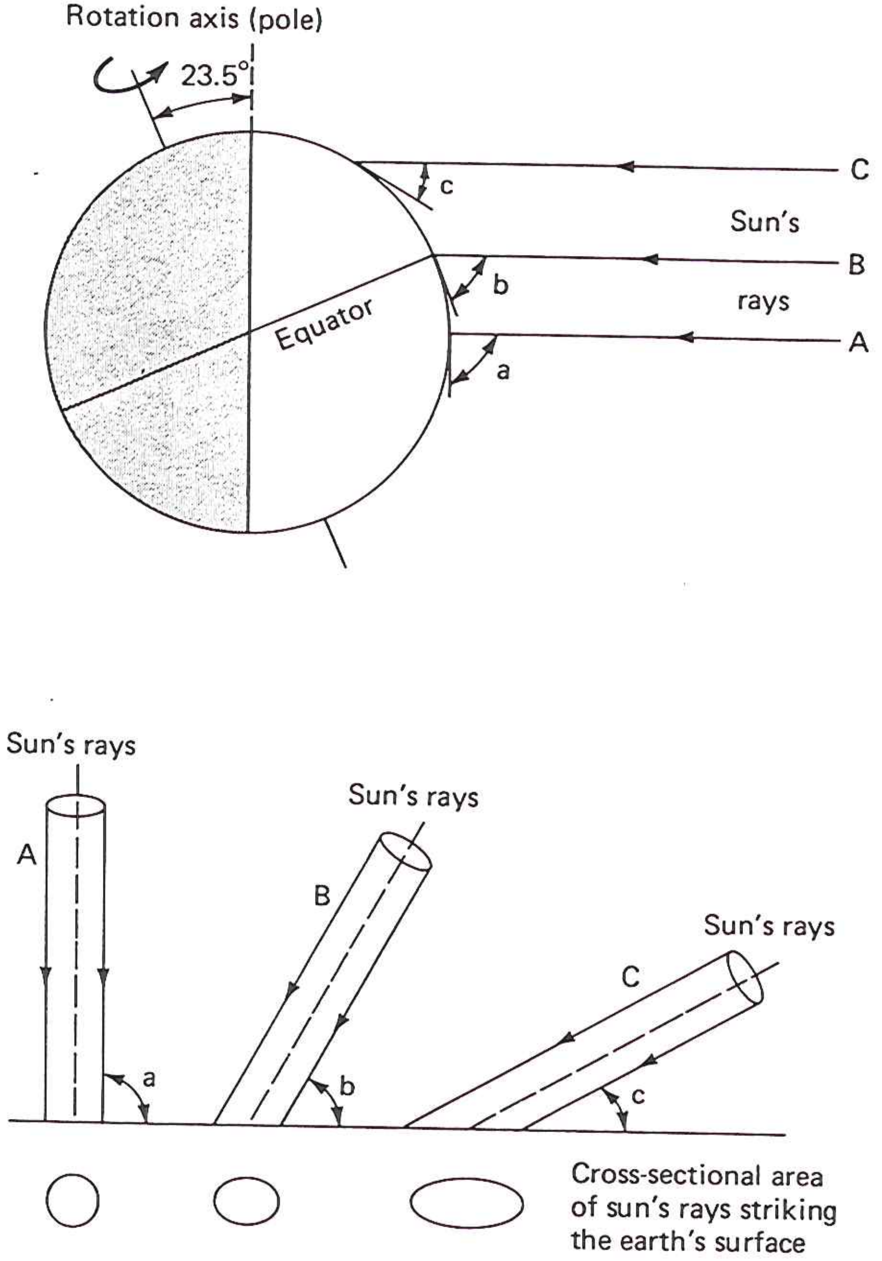


Figure 2.6 Schematic diagrams showing the variations of solar intensity (energy per unit area) with angle of incidence to the earth's surface. Lower angles (higher latitudes) result in the same energy spread out over a larger area and, thus, in a lower intensity of radiation. Scene depicted is for Northern 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.)

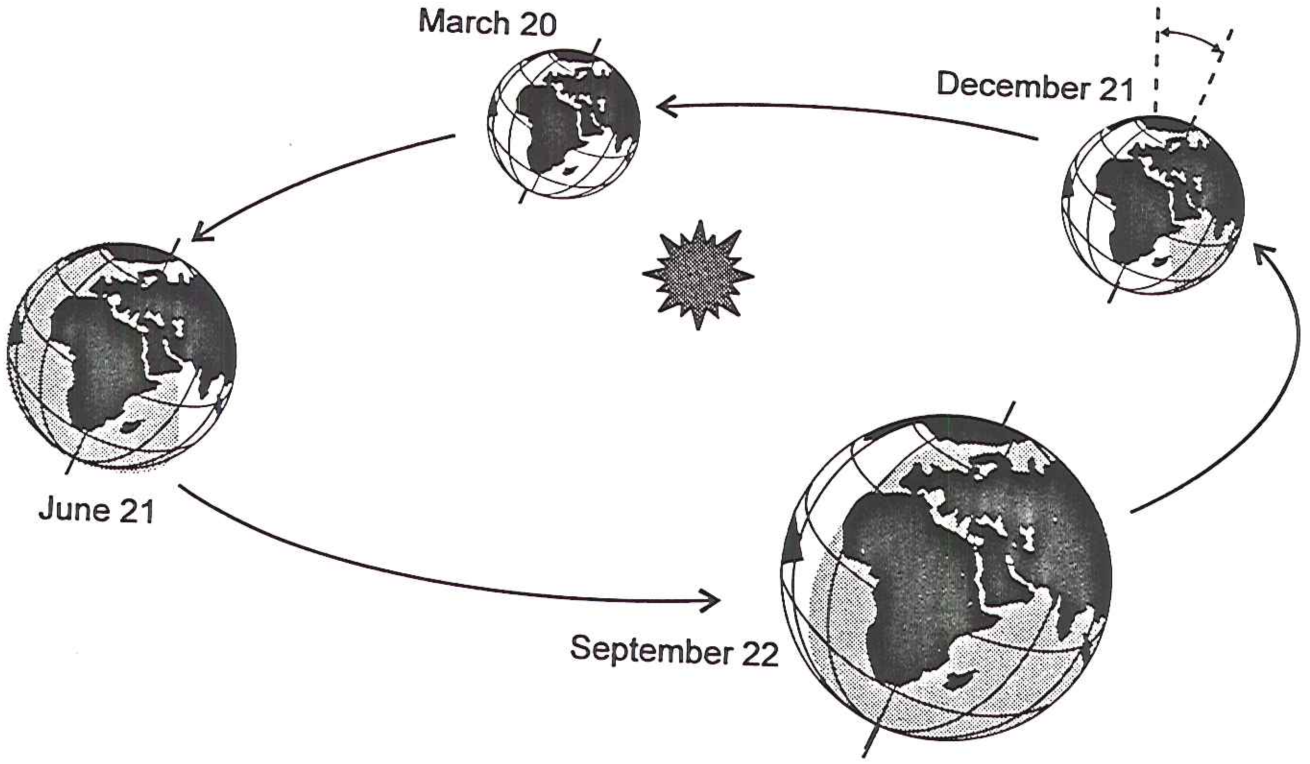
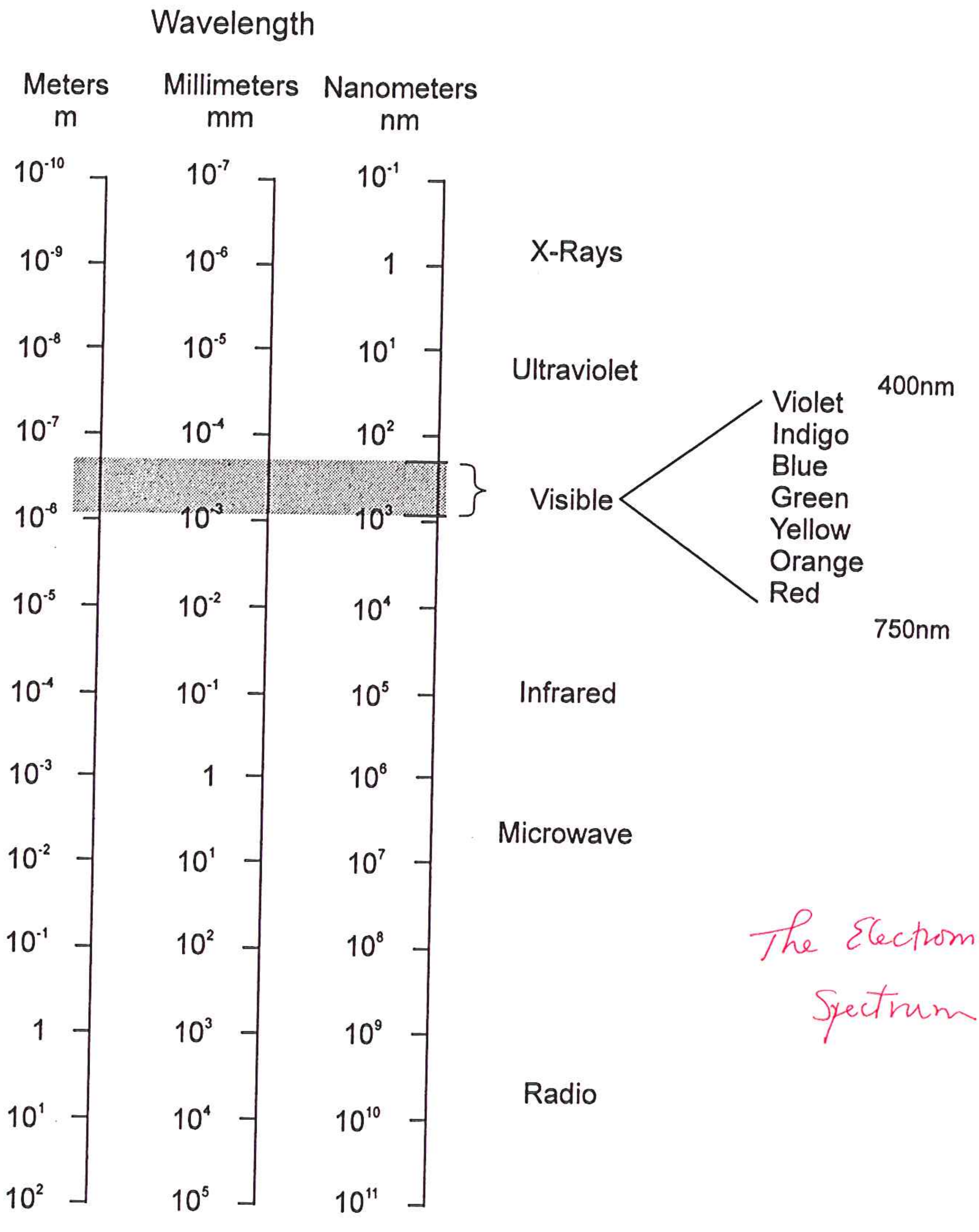


Figure 3.11. March of the seasons; as the tilted Earth revolves around the Sun, changes in the distribution of sunlight cause the succession of the seasons. (Imbrie and Imbrie 1986)



The Electromagnetic Spectrum

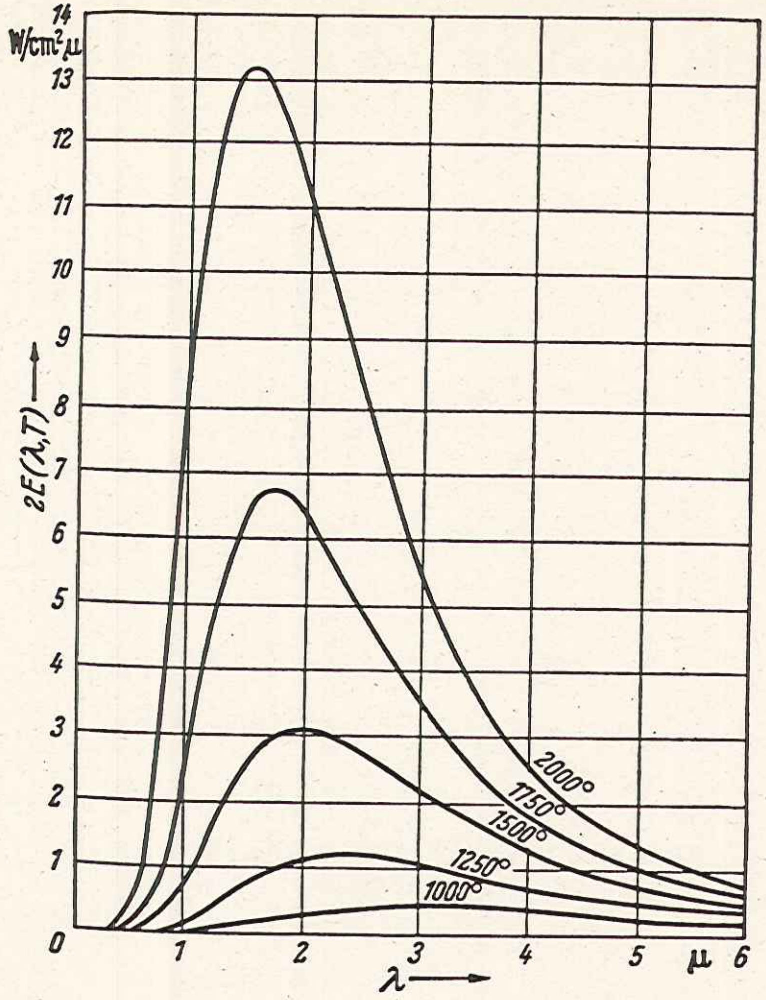


Figure 6.5. Power per unit wavelength per steradian emitted by a blackbody emitter at different temperatures. (Finkelnburg, 1964, p. 45, Fig. 20.)

Planck Distributio

Energy emitted per meter^2 with wavelength between λ and $\lambda + d\lambda$:

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

$c =$ 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

<u>Depends only upon the temperature</u> !

peak at 500 nanometers
(why the sun is yellow)

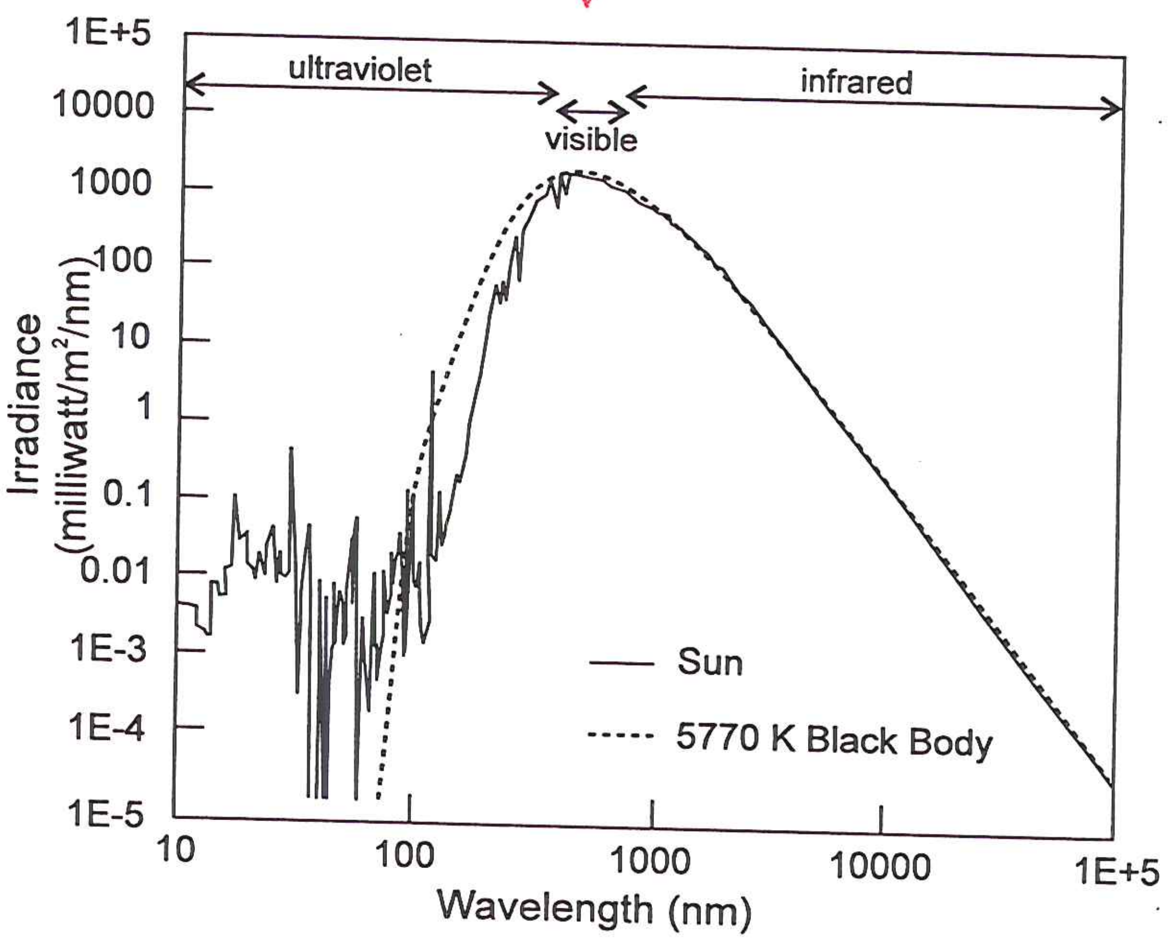


Figure 2.6. The Sun's spectral irradiance typical of solar minimum conditions compared with the spectrum of a blackbody radiator at 5,770 K (Lean 1991). UV = ultraviolet; VIS = visible; IR = infrared.

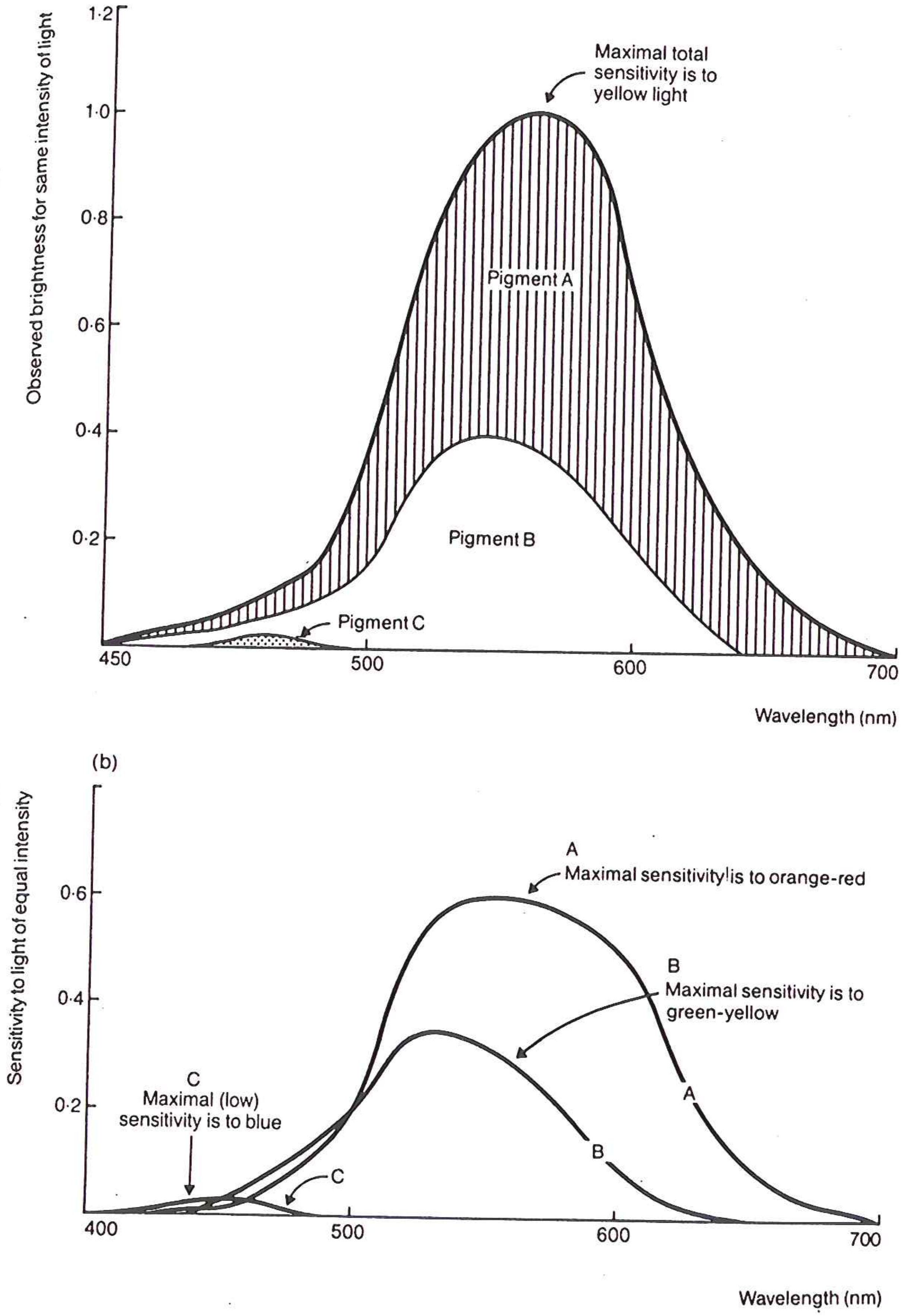


Figure 47. Brightness by day *(a) The full line shows the combined response of the cone cells to light of the same intensity but of different wavelengths. The approximate contributions of the three systems of cone pigments are indicated by the differently shaded areas. (b) The approximate sensitivities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

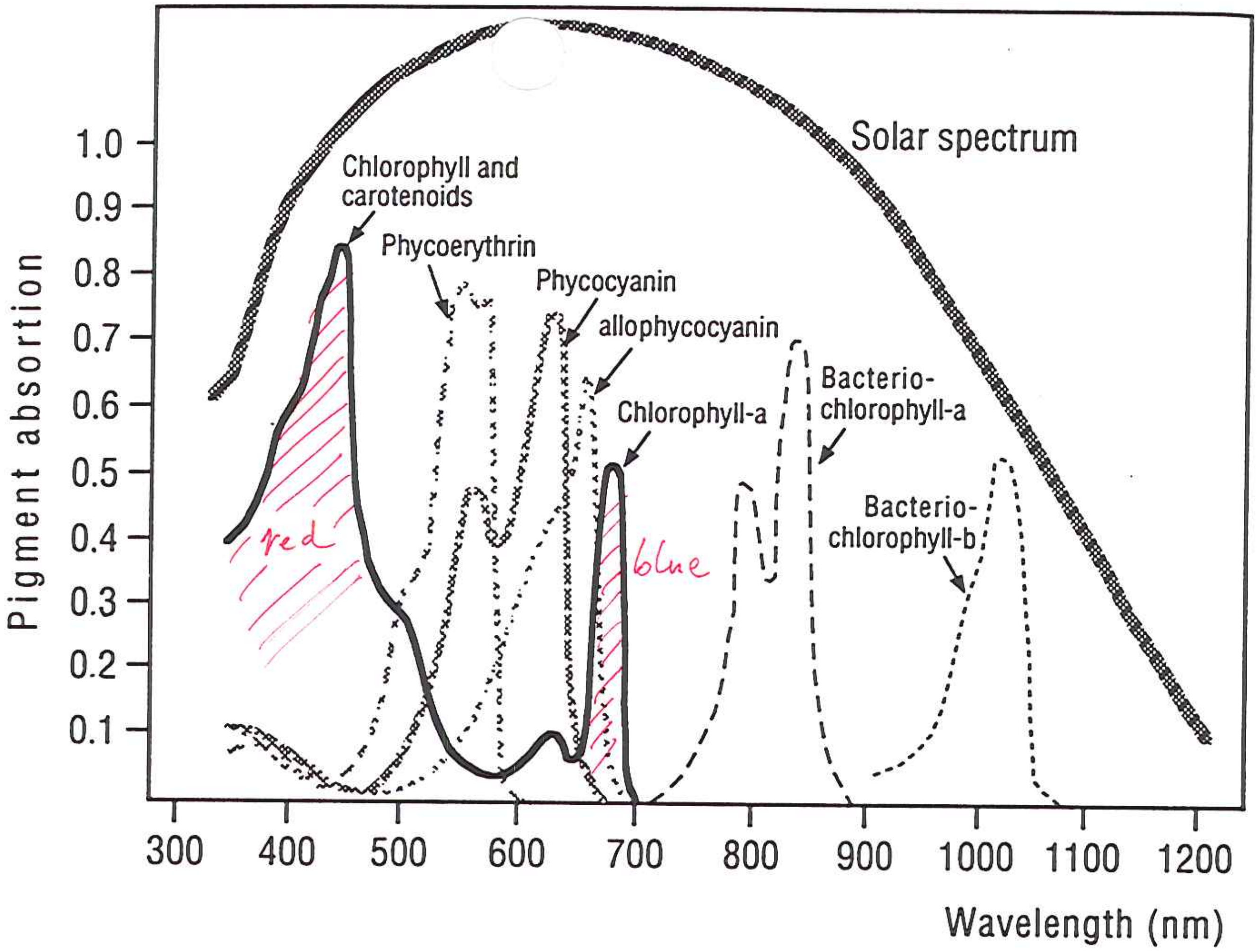
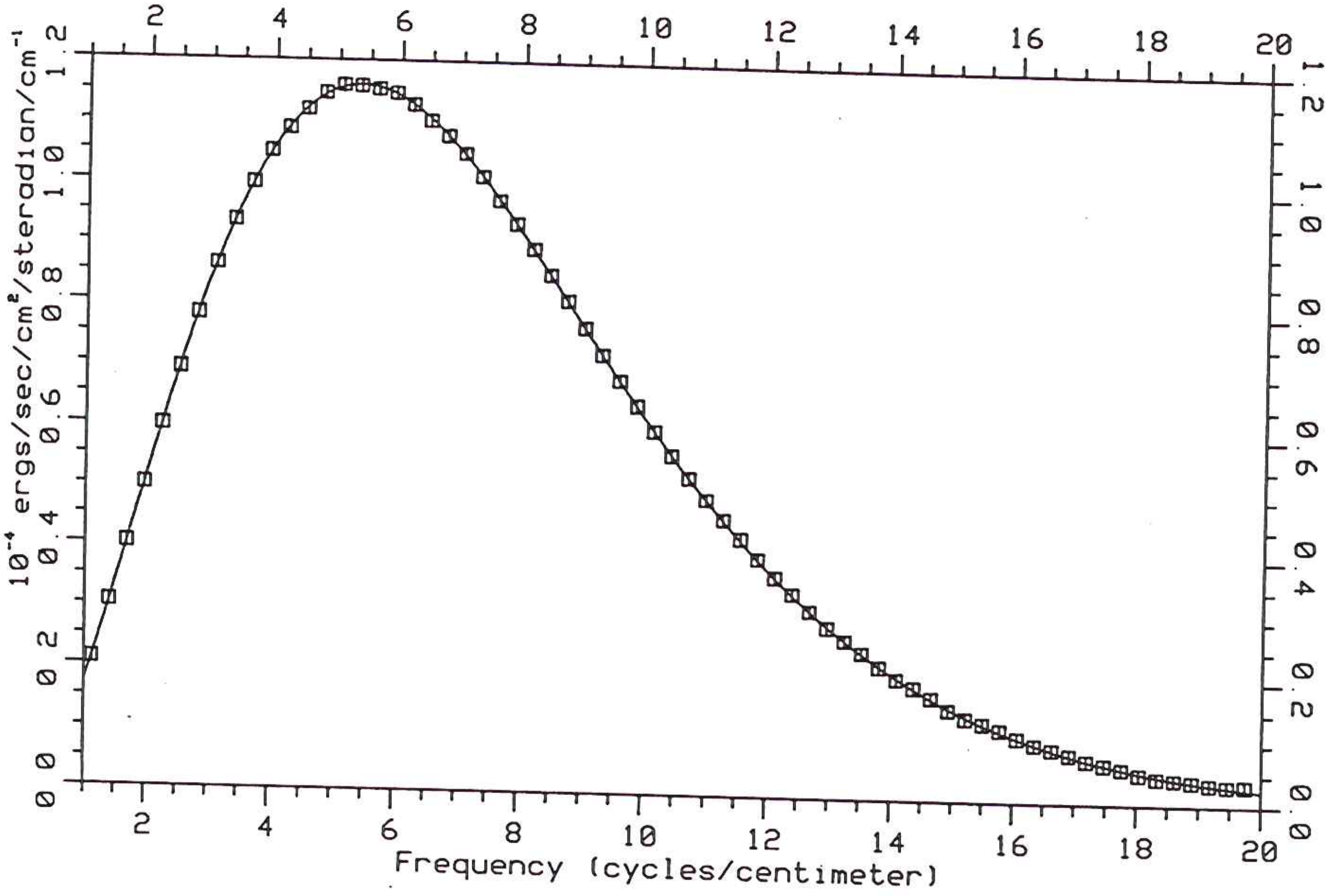


Figure 2.1 The solar spectrum as it reaches the surface of the earth. It overlaps with the important pigments of a variety of photosynthetic organisms. Chlorophyll *a*, the major pigment of higher plants, algae and cyanobacteria, absorbs red and blue light. In combination with carotenoids (such as β-carotene) they provide plants with their typical green colour. The major photosynthetic organisms of the world oceans, the cyanobacteria, absorb sunlight by a specialized pigment containing protein, the phycobilisome, which contains as major pigments phycoerythrin (±580 nm), phycocyanin (±620 nm) and allophycocyanin (±650 nm). All the light energy absorbed by the pigments can be used for photosynthesis. Bacteriochlorophyll *a* and bacteriochlorophyll *b* are the major pigments of two classes of photosynthetic bacteria absorbing in the near infrared part of the spectrum



10⁻⁴ ergs/sec/cm²/steradian/cm⁻¹
0.0
0.2
0.4
0.6
0.8
1.0
1.2
2
4
6
8
10
12
14
16
18
20
Frequency (cycles/centimeter)

Atmospheric composition

Major Components (concentration in percent by volume in dry air)	
Nitrogen, N_2	78.08
Oxygen, O_2	20.95
Argon, Ar	0.93 ← from decay of K
Minor Components (concentration in parts per million by volume, ppmv)	
Water vapor, H_2O	40–40,000 ← big range
* 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.

* important greenhouse gases

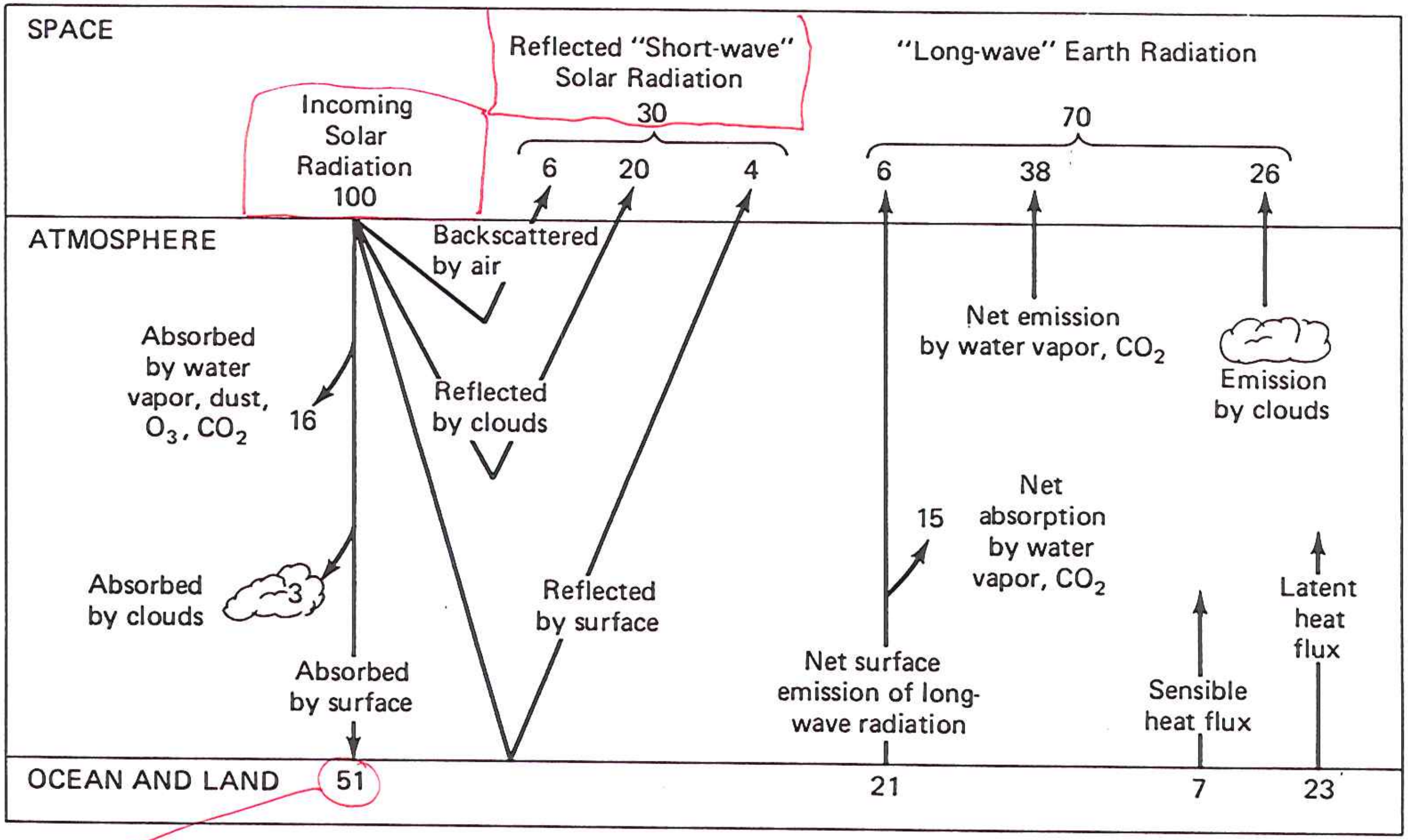


Figure 2.4 The mean annual radiation and heat balance of the atmosphere and earth. Units are assigned so that incoming solar radiation (0.5 cal/cm²/min) is set equal to 100. ("Short-wave" solar radiation is that with <4μm wave length; "long-wave" earth radiation is >4μm). (Adapted from U.S. Committee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

only half
of short-wavelength
radiation reaches
the ground

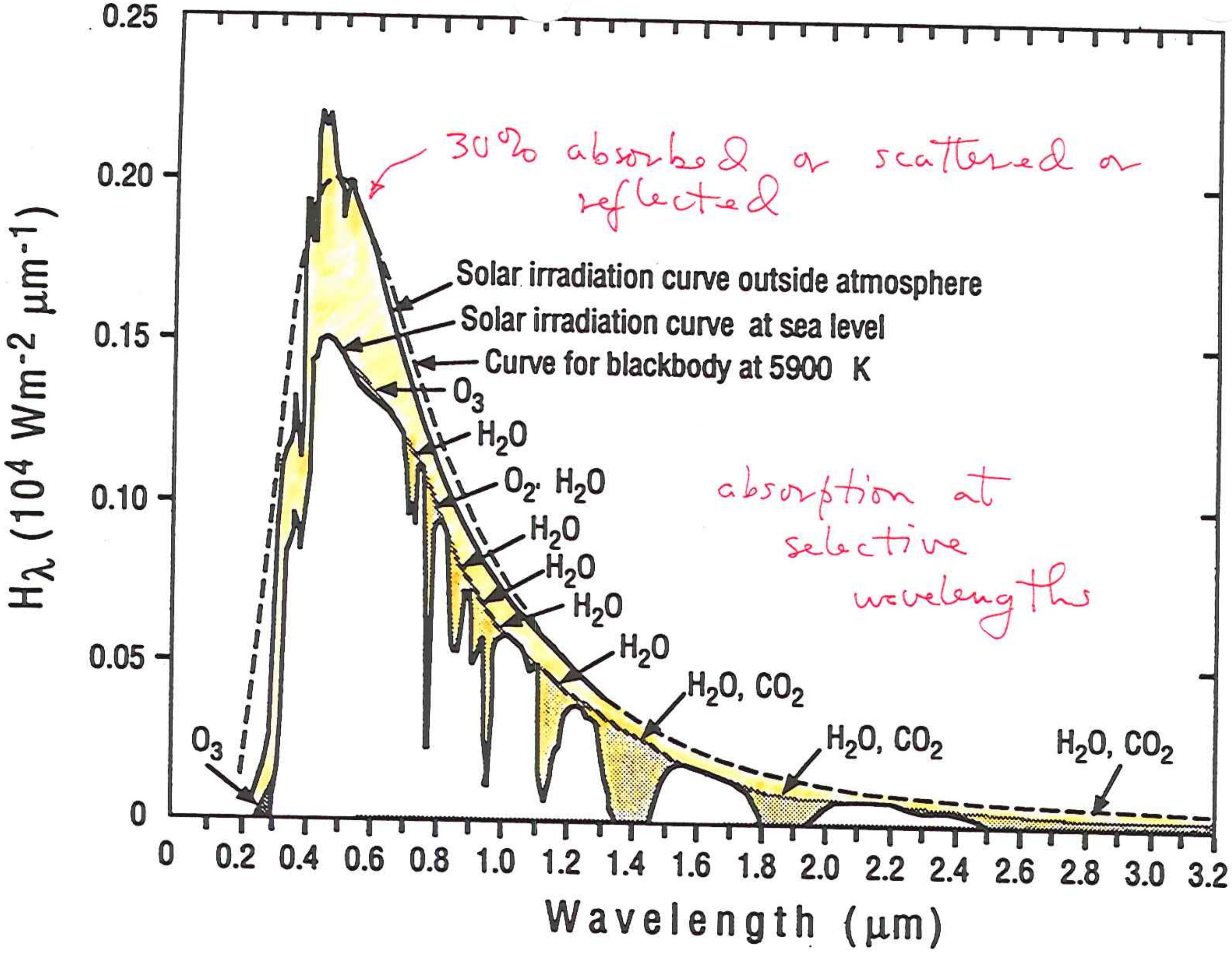
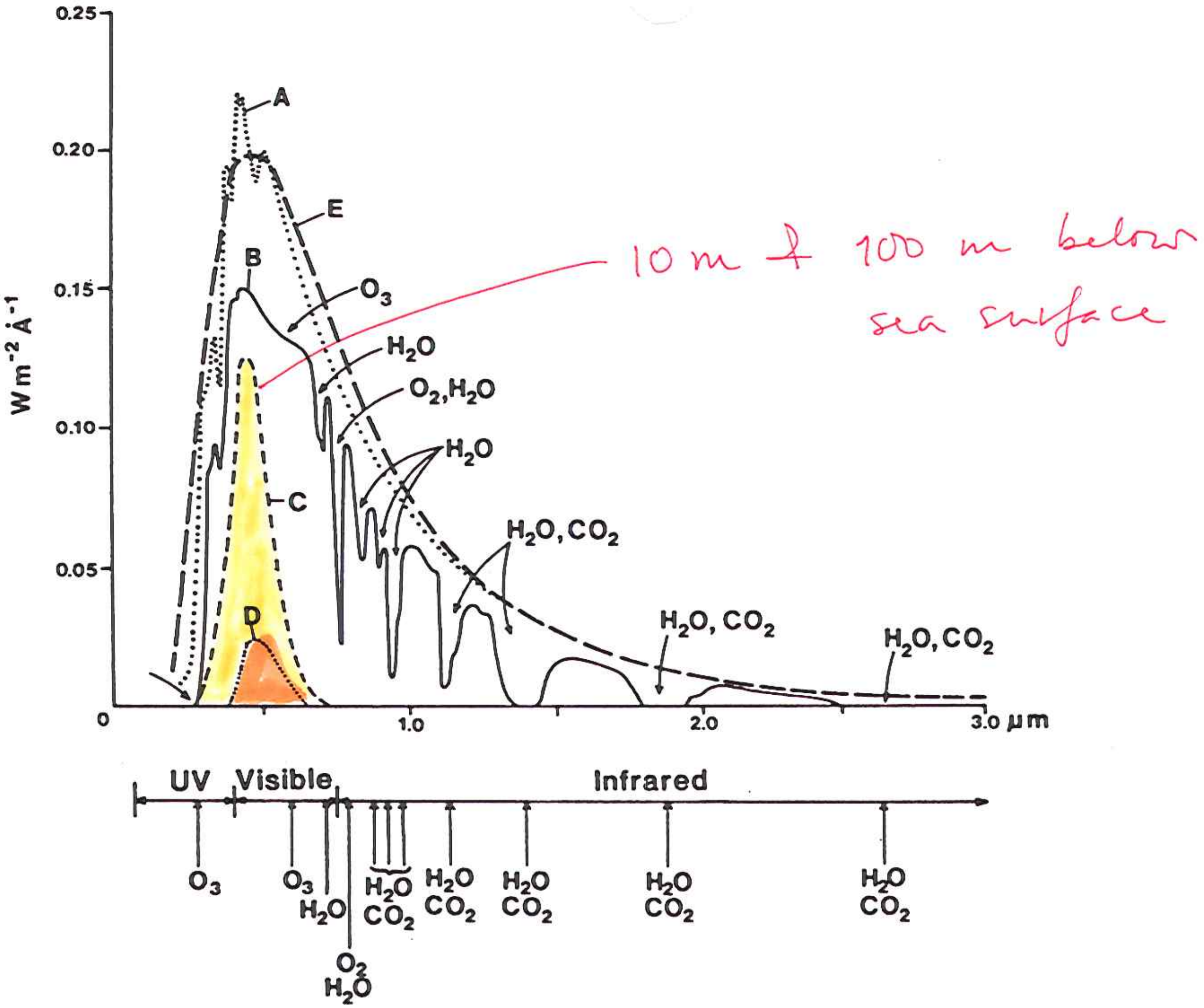


Figure 2.2 Spectral distribution of incident solar radiation outside the atmosphere and at sea level. Major absorption bands of some of the important atmospheric gases are indicated. (Reproduced by permission of McGraw-Hill from S. L. Valley (ed.), *Handbook of Geophysics and Space Environments*, McGraw-Hill,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



10 m & 100 m below sea surface

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.)

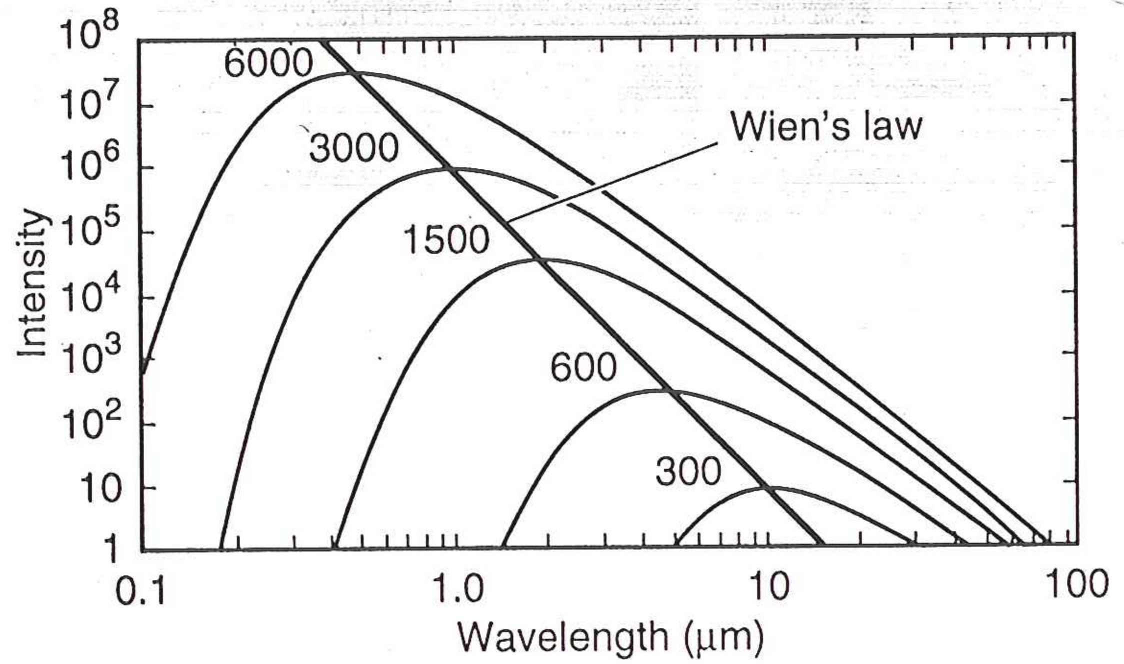
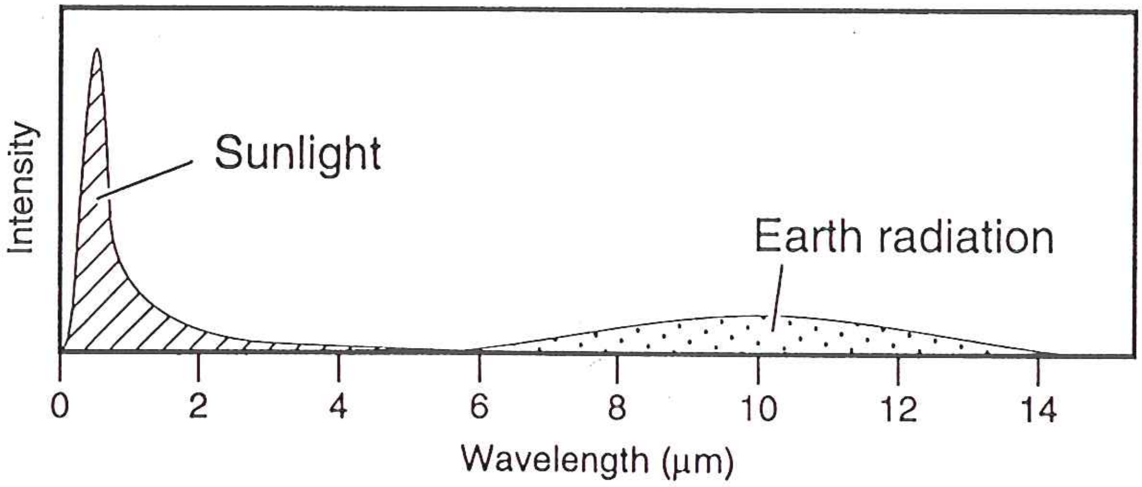


Figure 3.7 Blackbody radiation spectra as a function of temperature (kelvin), over the entire range of temperatures relevant to environmental studies. The values are displayed here on a log-log graph, so that both the wavelength and intensity scales are greatly compressed and cover many orders of magnitude.

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.



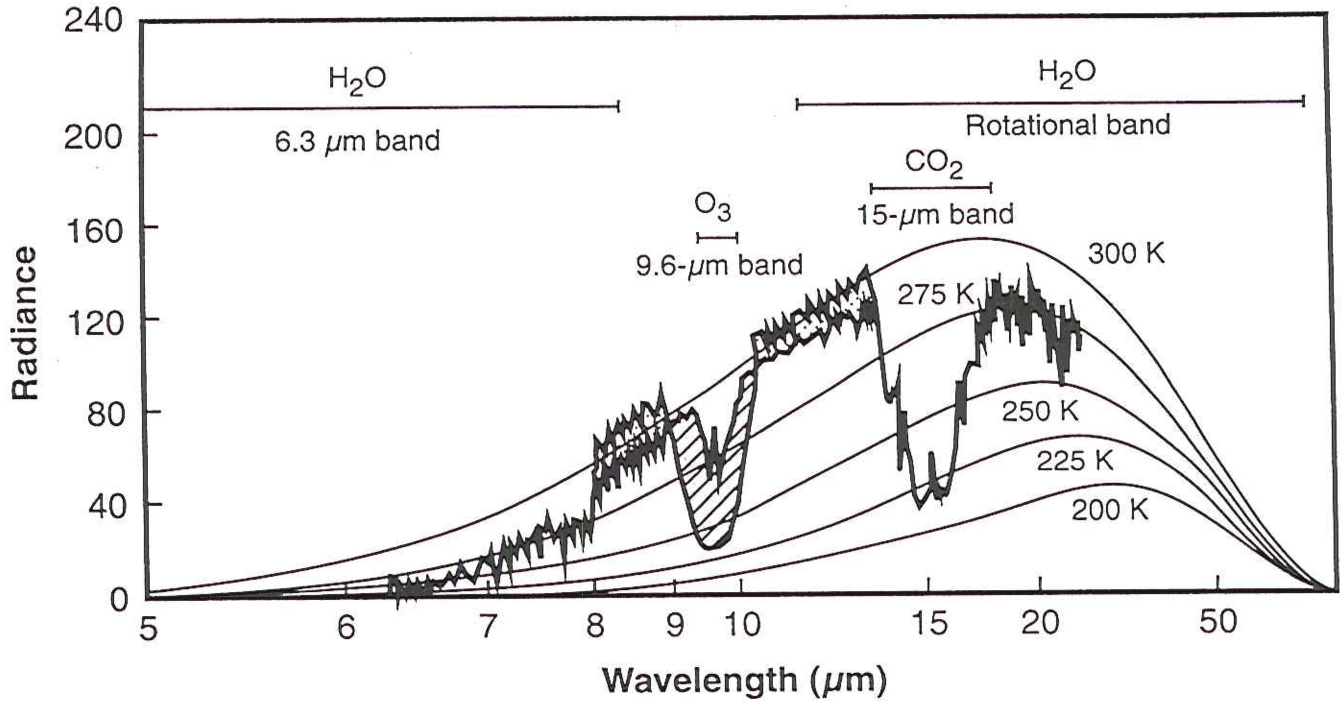
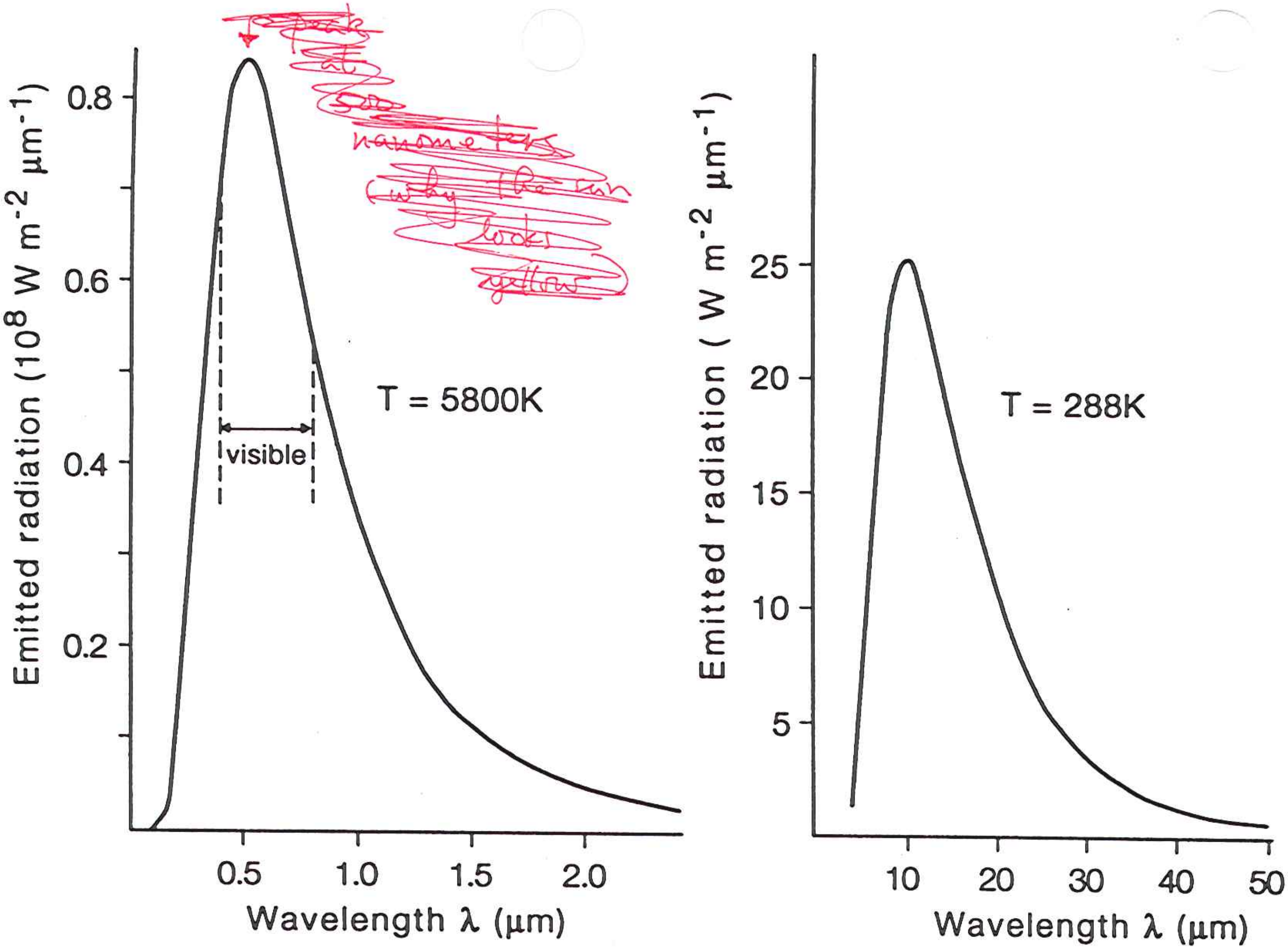


Figure 11.11 The spectrum of radiation emitted by the Earth to space, as measured from a satellite. The spectrum is the jagged line, which defines the intensity of the radiation as a function of wavelength. Several ideal blackbody emission curves (Planck functions) corresponding to temperatures ranging 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.)



[redacted]

Figure 1.2 The emission spectra of a black body of 5 800 K such as the sun on the left and of a black body of 288 K such as the earth surface on the right. Note the differences in scale of both figures. Even at the peak of the earth's emission spectrum the solar emission is much higher.

NB — this is
W/m^2 — not
W/m^2 per μm
That's why the
peak is at
18 μm
instead of 10 μm.

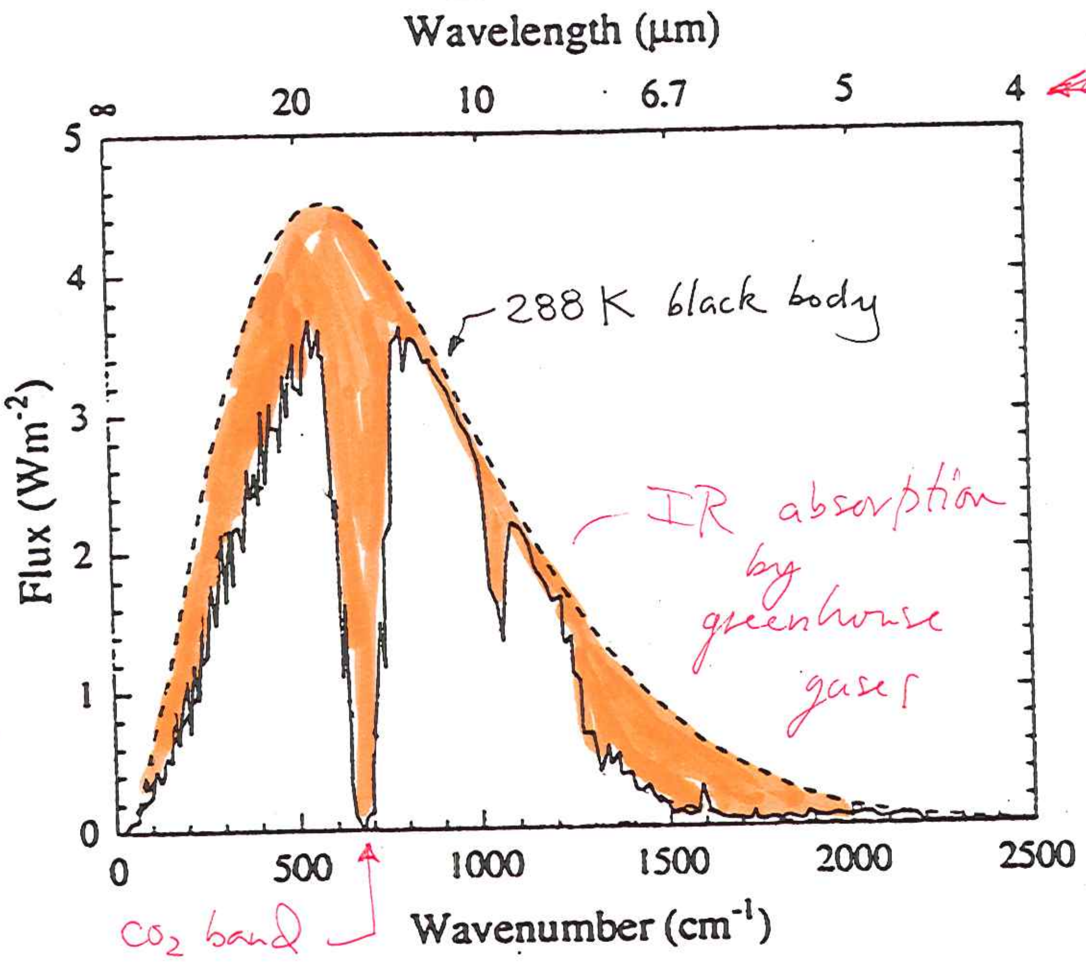


Fig. 1. Outgoing clear-sky terrestrial radiation calculated by K.P. Shine for a region where the surface temperature is 294 K (Shine et al. 1990). The dashed lines show the hypothetical energy flux which would be observed if none of the outgoing radiation were intercepted in the atmosphere.

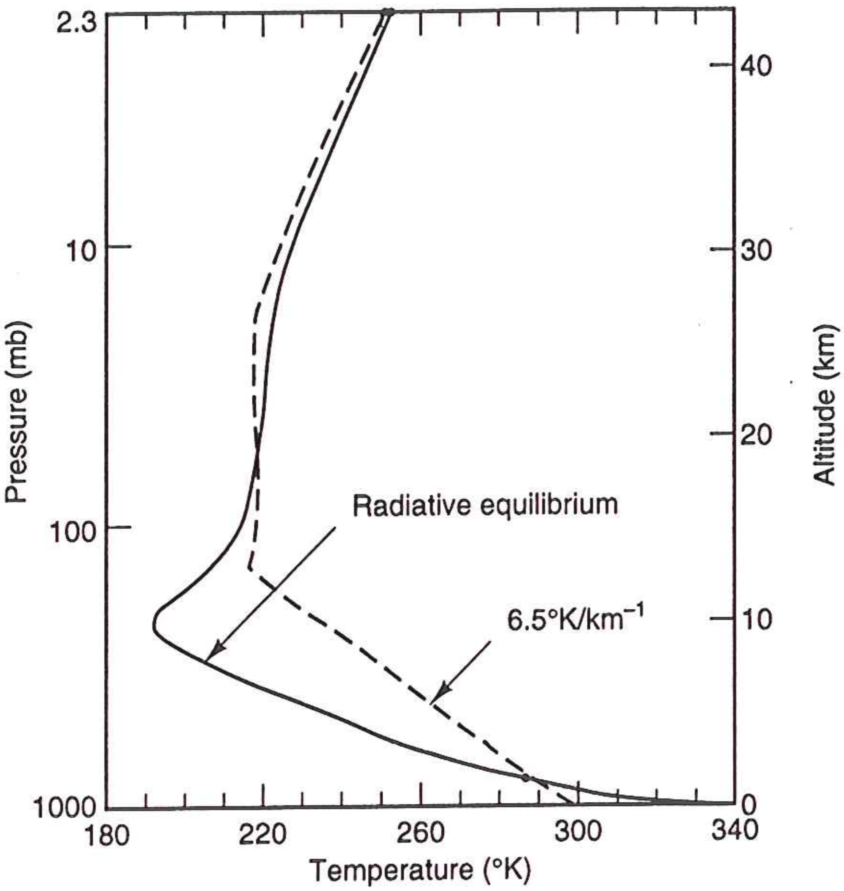


FIGURE 3–6 The variation of temperature with altitude in a radiative equilibrium model assuming: (1) pure radiative equilibrium only, and (2) convection and precipitation occurring, resulting in a lapse rate (or decrease in temperature with increasing altitude) of 6.5°K km^{-1}. (After S. Manabe and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

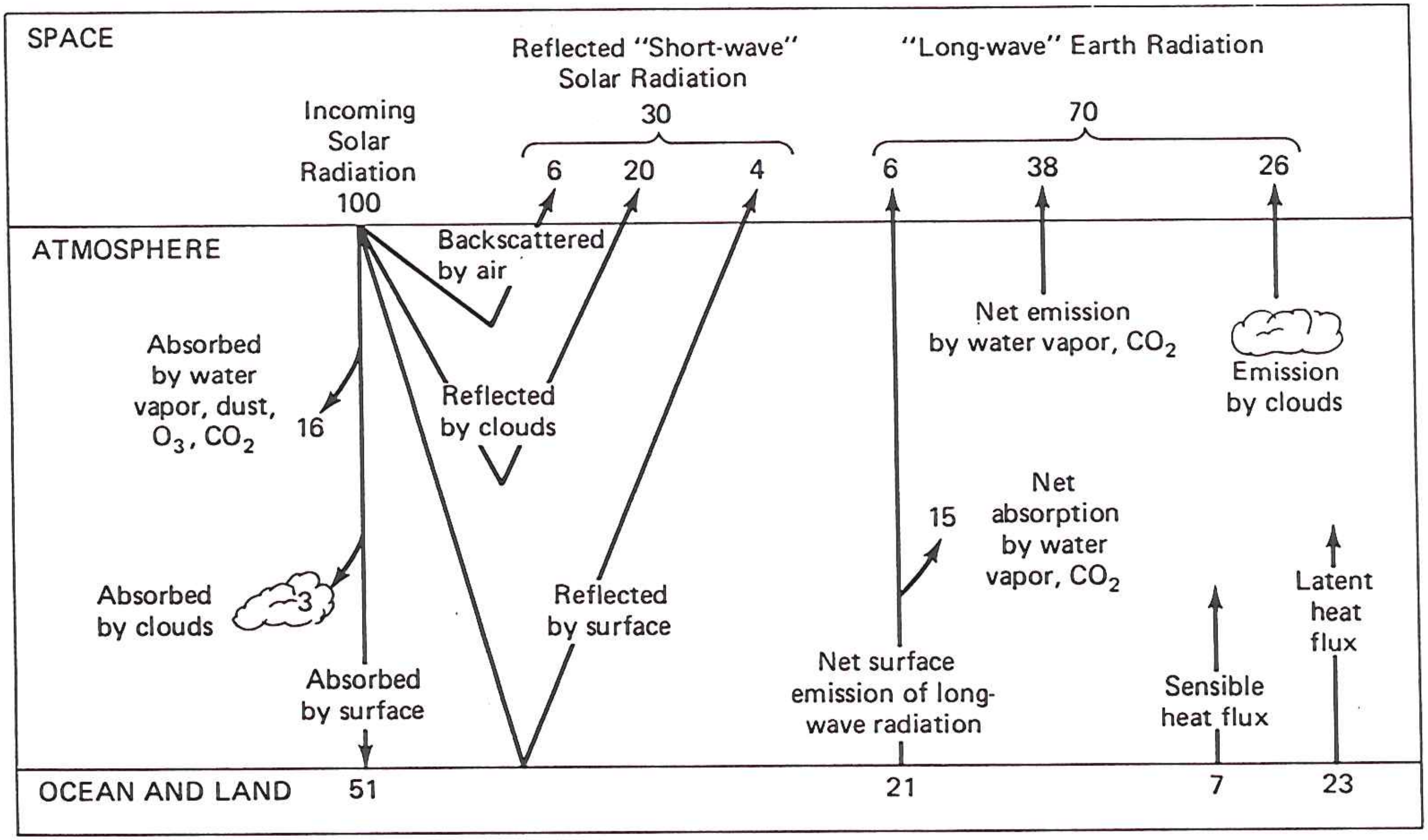


Figure 2.4 The mean annual radiation and heat balance of the atmosphere and earth. Units are assigned so that incoming solar radiation (0.5 cal/cm²/min) is set equal to 100. ("Short-wave" solar radiation is that with <4μm wave length; "long-wave" earth radiation is >4μm). (Adapted from U.S. Committee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

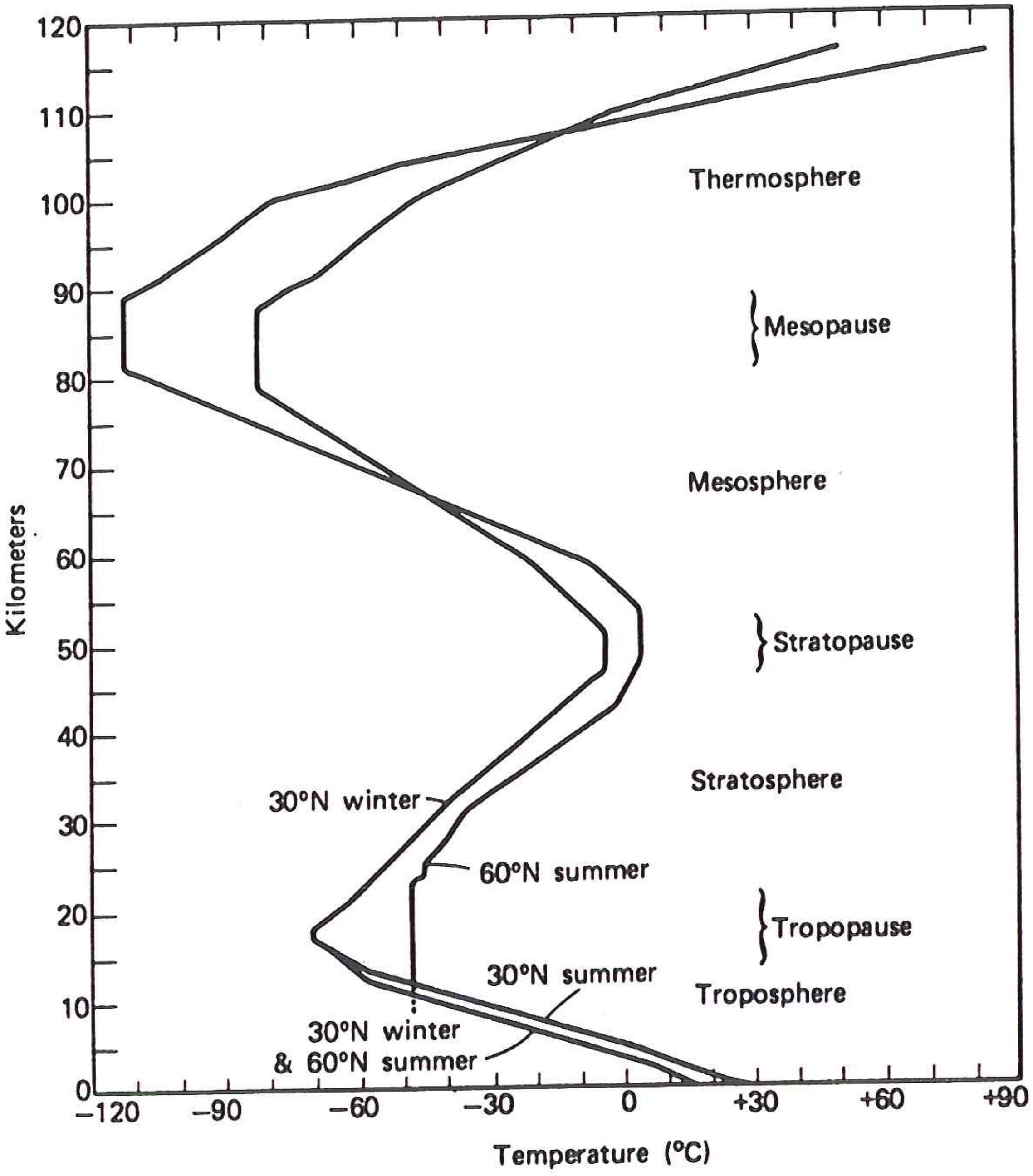
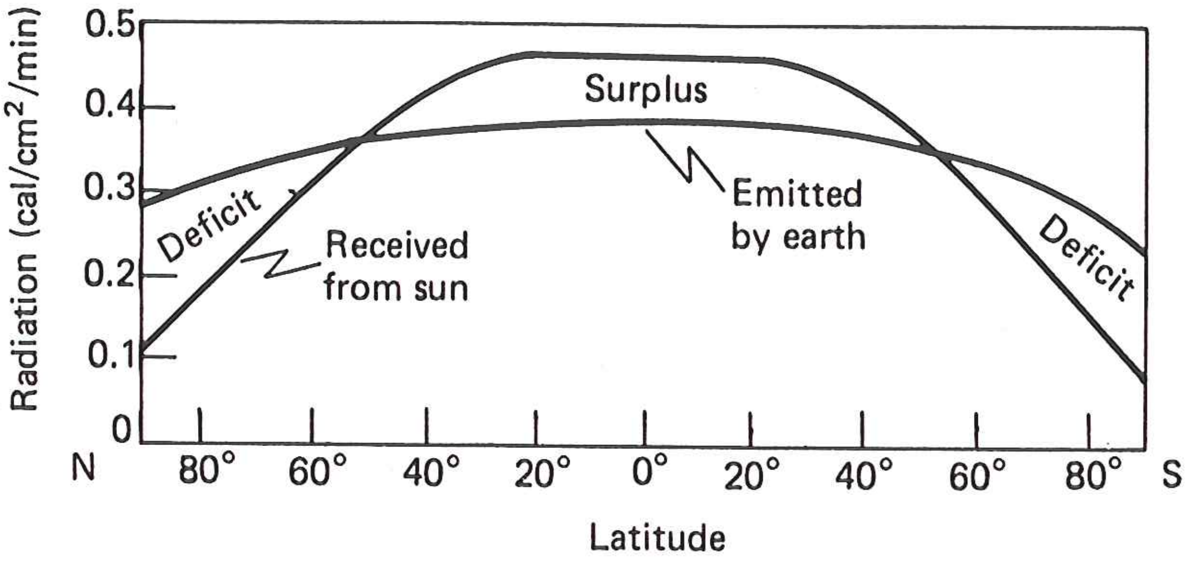
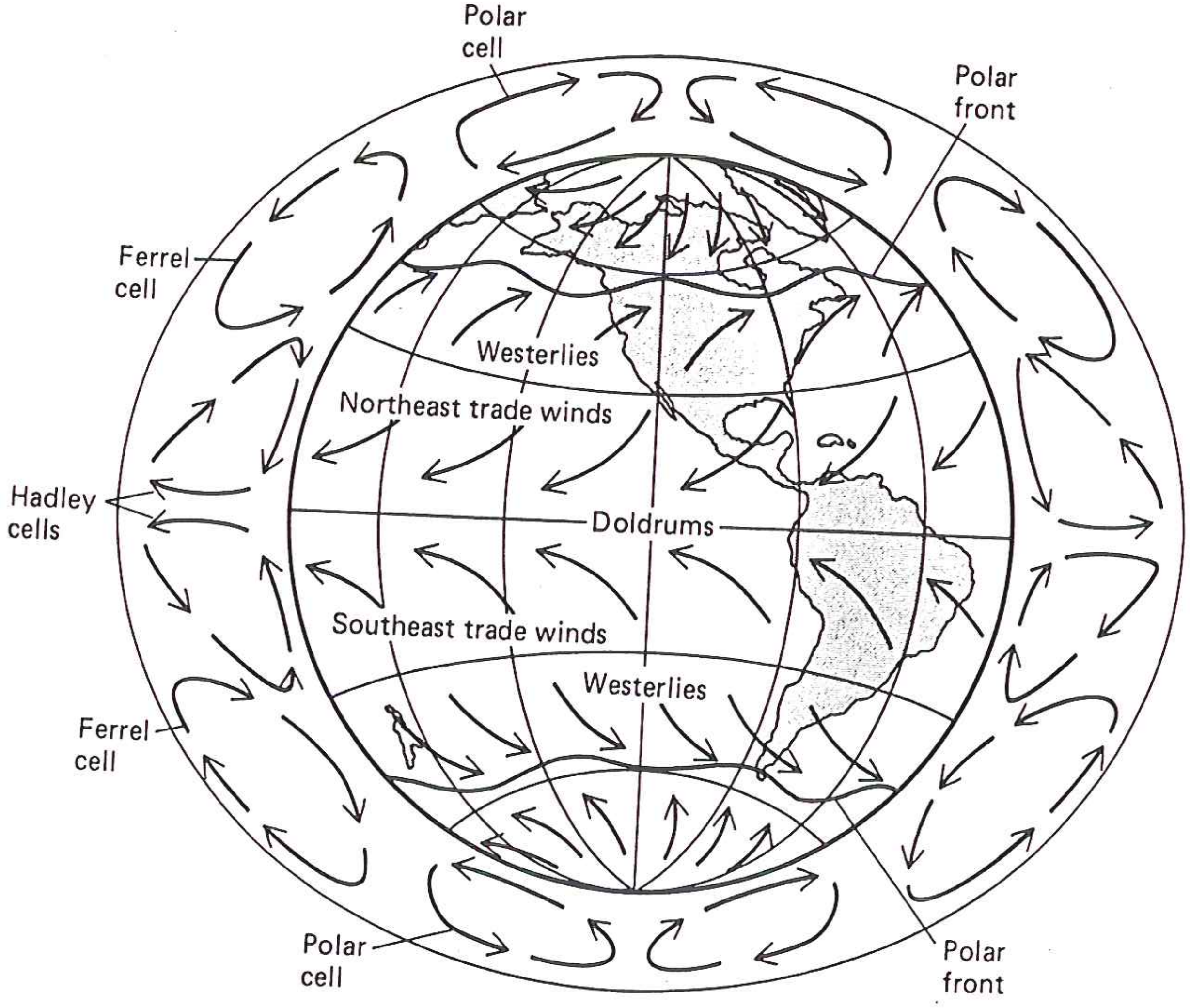


Figure 13.5. Thermal structure of the atmosphere, 0–120 km. (Emiliani, 1987, p. 16.)


Radiation (cal/cm²/min)
0.5
0.4
0.3
0.2
0.1
0
N
80°
60°
40°
20°
0°
20°
40°
60°
80°
S
Latitude
Surplus
Deficit
Deficit
Received from sun
Emitted by earth


Polar cell
Polar front
Ferrel cell
Westerlies
Northeast trade winds
Hadley cells
Doldrums
Southeast trade winds
Westerlies
Ferrel cell
Polar cell
Polar front

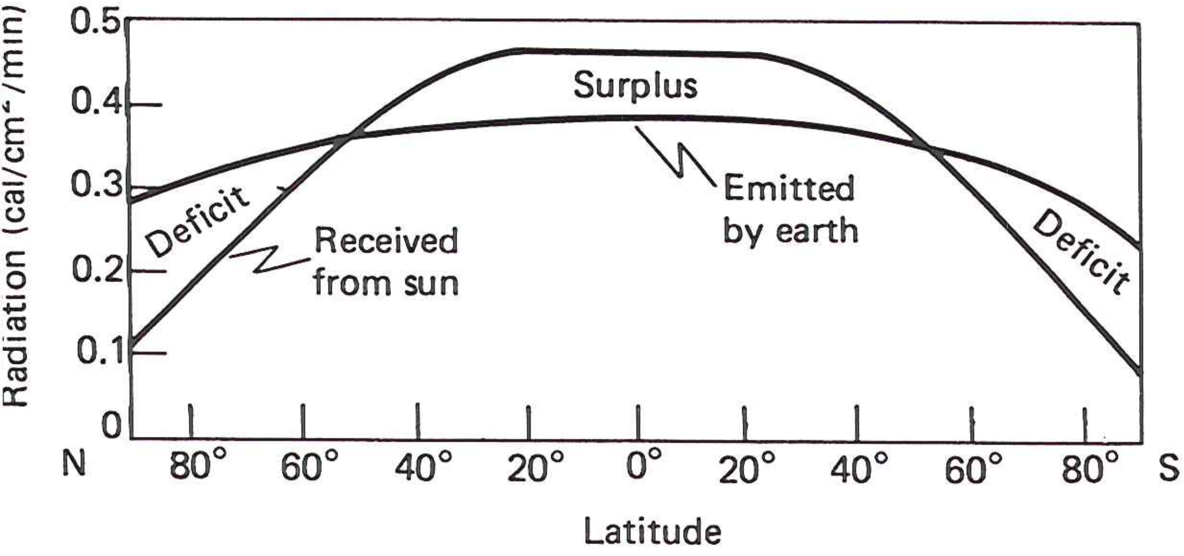


Figure 2.8 Mean annual radiation absorbed from the sun and radiated from the earth to space, as a function of latitude. (Adapted from A. Miller, et al. *Elements of Meteorology, 4th ed.* Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

The equatorial heat surplus drives the large-scale circulation of the atmosphere — responsible for prevailing weather patterns

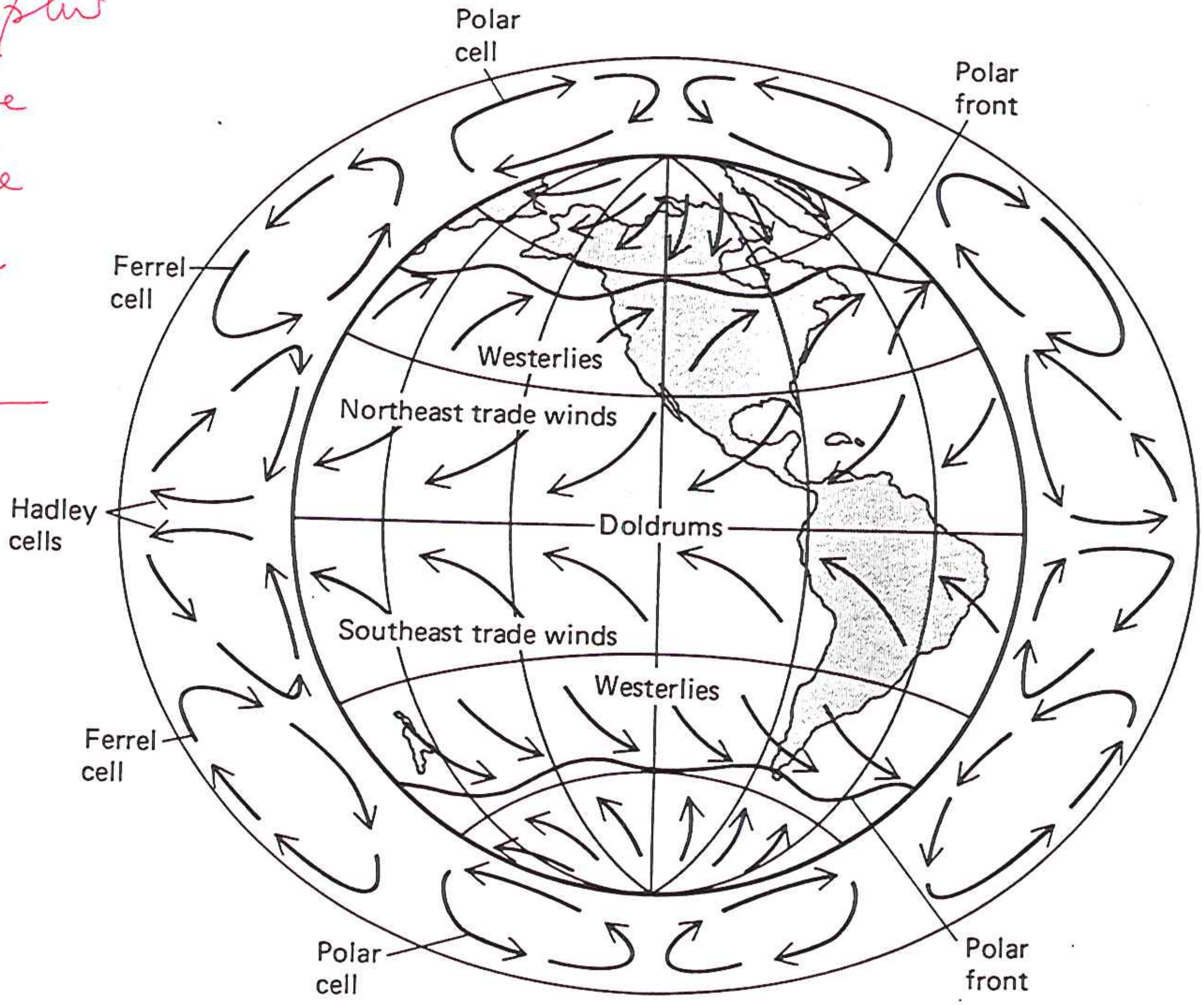


Figure 2.10 Schematic representation of the general circulation of the atmosphere. (Modified from A. Miller, et al. *Elements of Meteorology, 4th ed.* Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

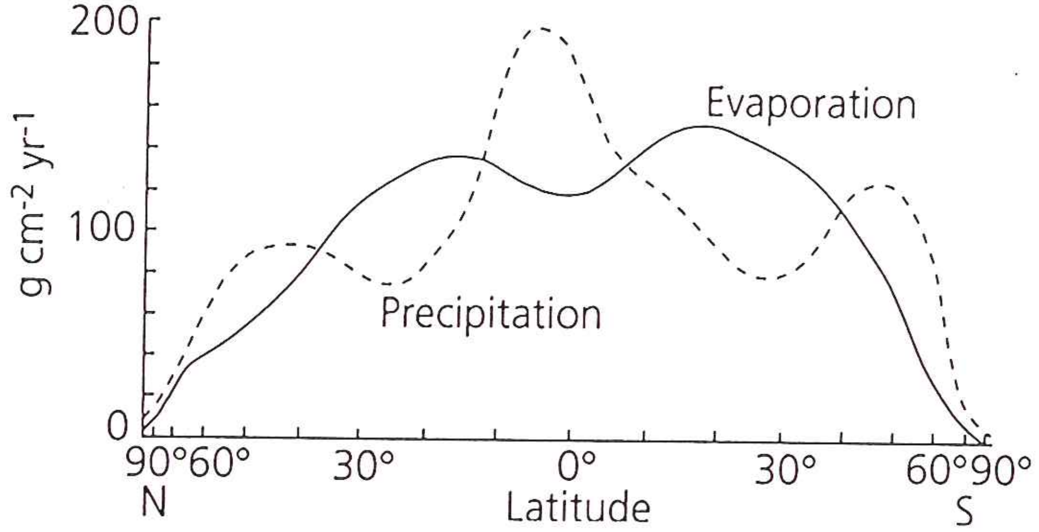


Fig. 1.19 Zonally averaged precipitation and evaporation as a function of latitude.

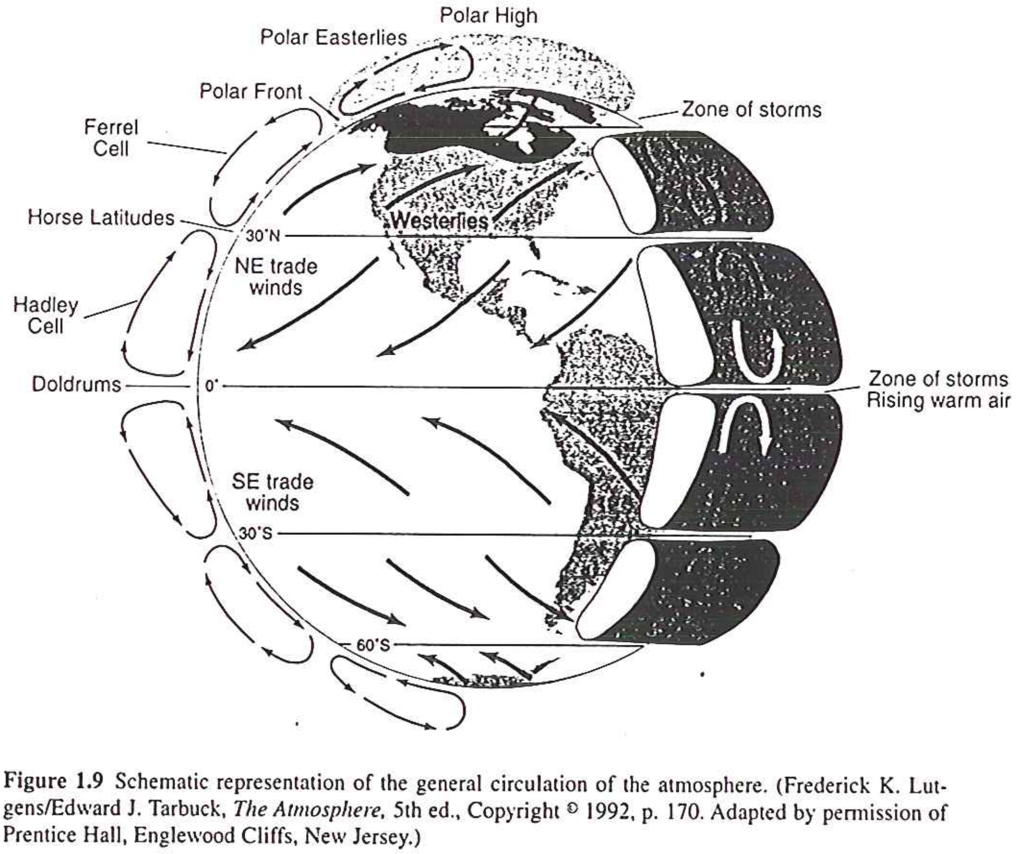


Figure 1.9 Schematic representation of the general circulation of the atmosphere. (Frederick K. Lutgens/Edward J. Tarbuck, *The Atmosphere,* 5th ed., Copyright © 1992, p. 170. Adapted by permission of Prentice Hall, Englewood Cliffs, New Jersey.)

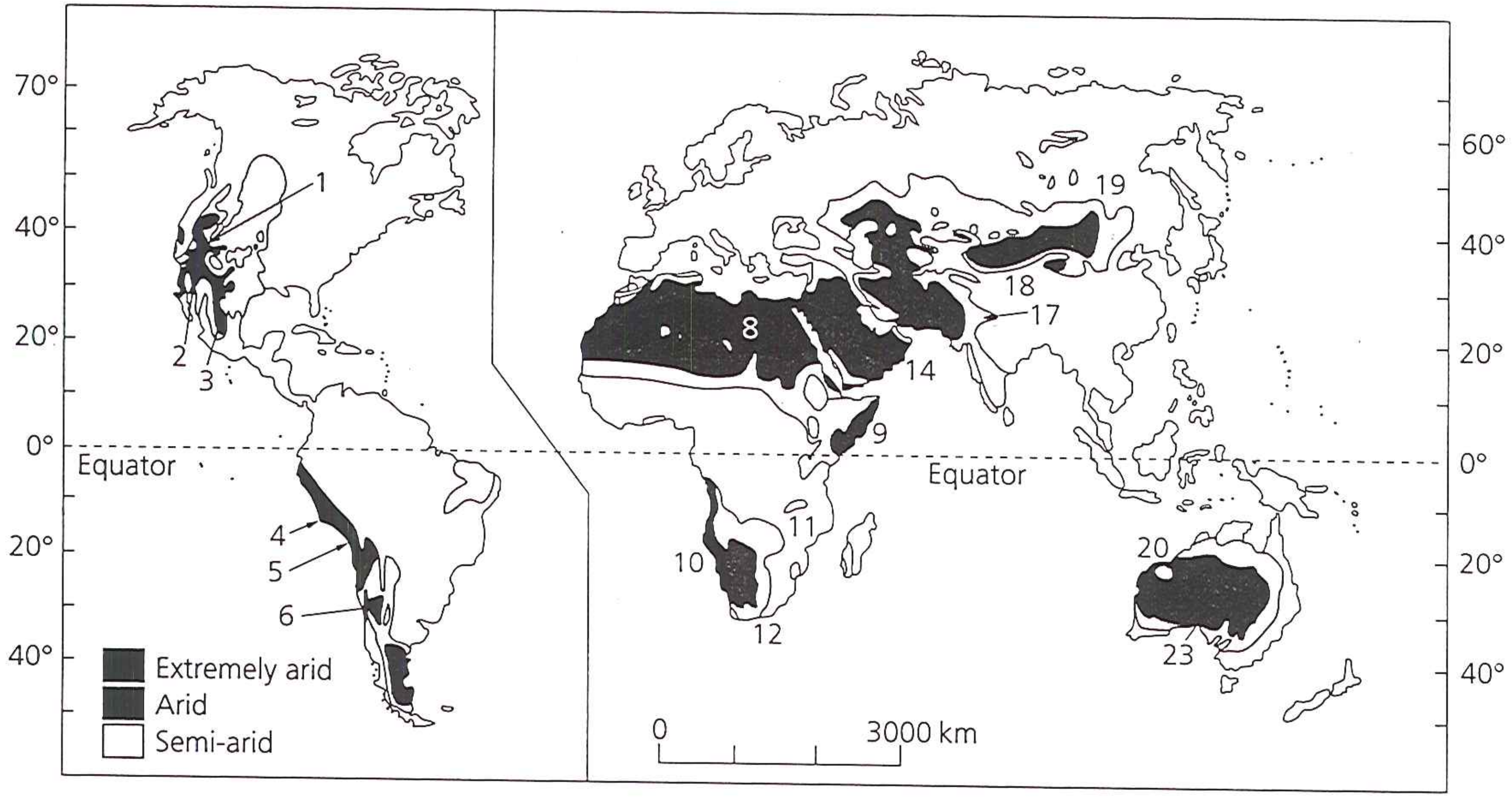


Fig. 10.24 Distribution of the world's drylands. The main deserts are: (1) Great Basin; (2) Sonoran; (3) Chihuahuan; (4) Peruvian; (5) Atacama; (6) Monte; (7) Patagonian; (8) Sahara; (9) Somali-Chabli; (10) Namib; (11) Kalahari; (12) Karroo; (13) Arabian; (14) Rub al Khali; (15) Turkistan; (16) Iranian; (17) Thar; (18) Taklimakan; (19) Gobi; (20) Great Sandy; (21) Simpson; (22) Gibson; (23) Great Victoria; (24) Sturt. After Cooke & Warren (1973) [32] and Greeley & Iversen (1985) [9].

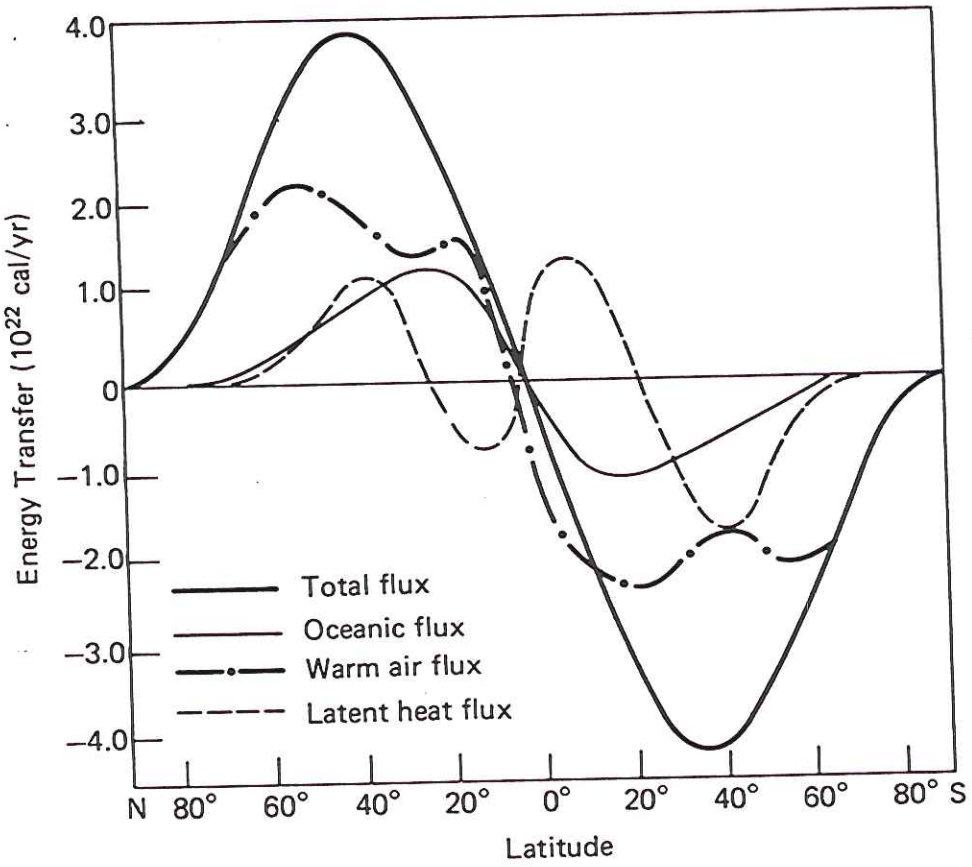


Figure 2.9 Rates of poleward energy transport by various mechanisms. Positive energy transfer values refer to northward transport and negative values to southward transport. (Adapted from W. D. Sellers, *Physical Climatology*. Copyright © 1965 by The University of Chicago Press. All rights reserved.)

The total poleward flux of energy is the sum of three components

ABLE 2.2 Fluxes in the Water Cyc with Methods of Determination

Process	Flux		Source
	km³/yr	cm/yr[a]	
Precipitation on land	110,300	74	Lvovitch 1973
Evaporation from land	72,900	49	(Precipitation on land minus runoff)
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)

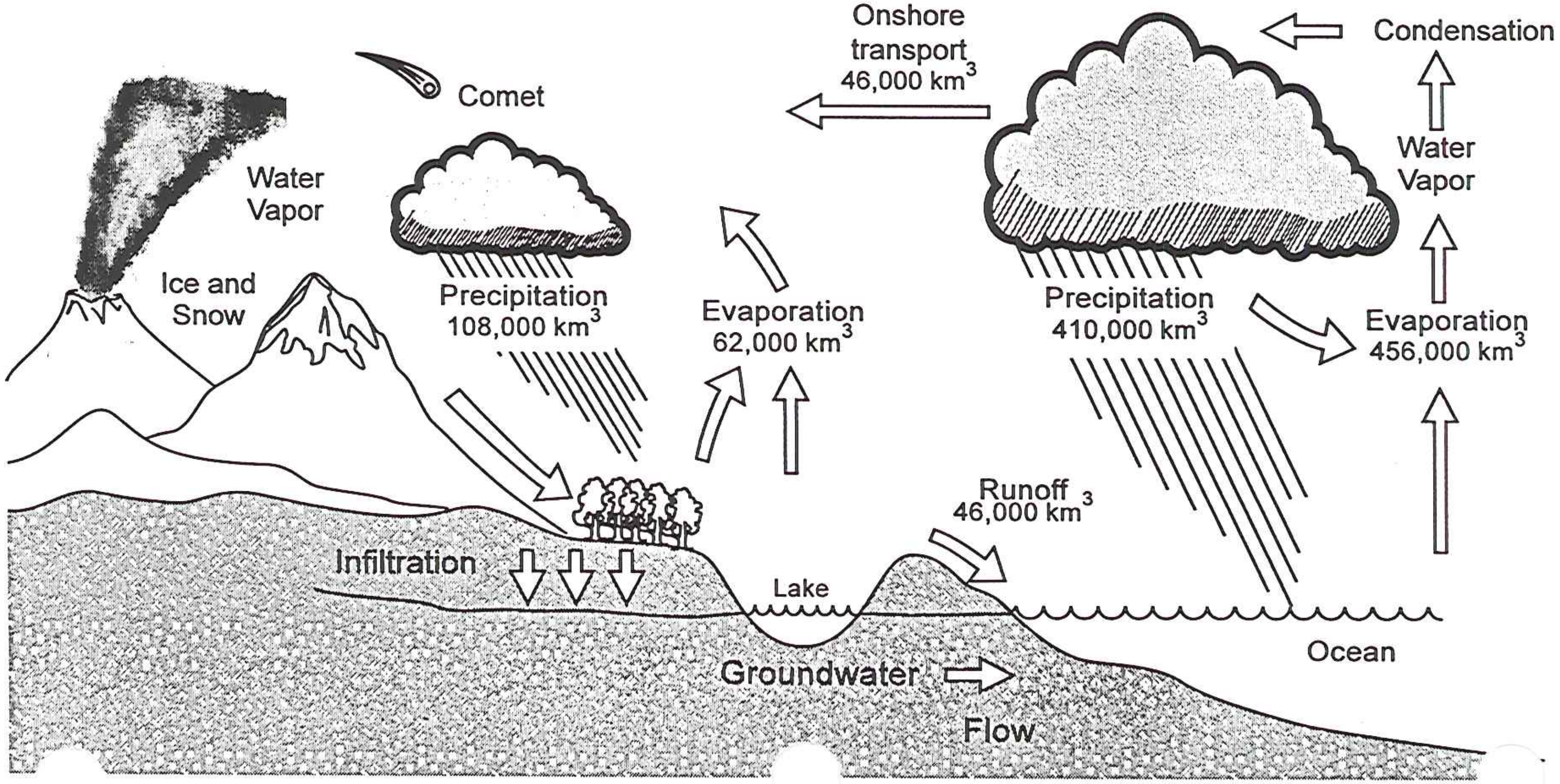
Note: Because of use of different areas, values in cm/yr do not balance between land and oceans.

[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).

The hydrologic cycle
moves water constantly
between aquatic,
atmospheric, and
terrestrial compartments
driven by solar energy
and gravity. Total annual
flows shown here are in
cubic kilometers.
(After Press and
Siever 1986)

These numbers seem a little off — don't agree with Berner's table

extra from here



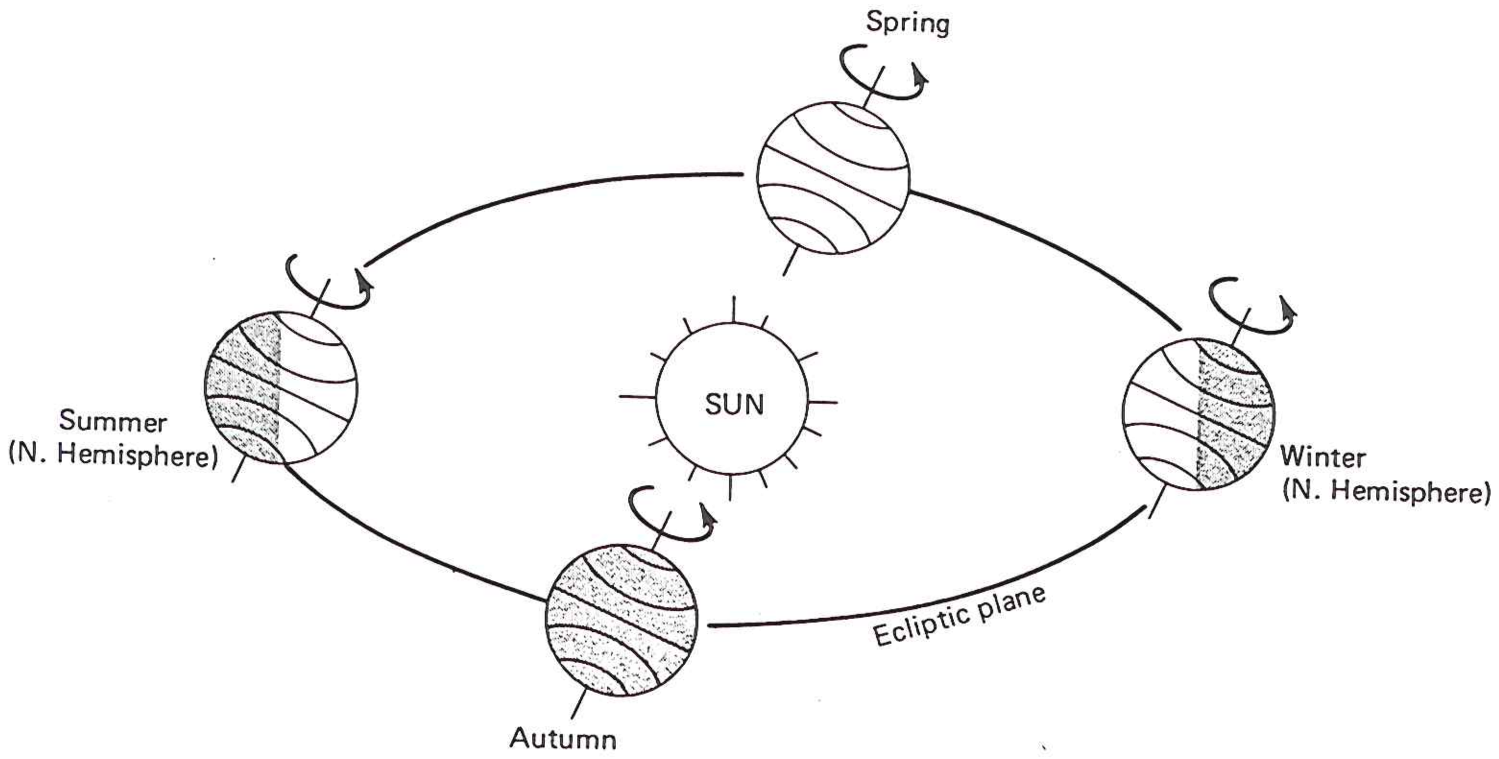


Figure 2.7 Revolution of the earth in its orbit around the sun, showing the changing seasons (also length of day). The seasons given are for the Northern Hemisphere; they are reversed in the Southern Hemisphere. (Modified from A. Miller, et al. *Elements of Meteorology, 4th ed*. Copyright © 1983 by Charles E. Merrill Publ. Co. Reprinted by permission of the publisher.)

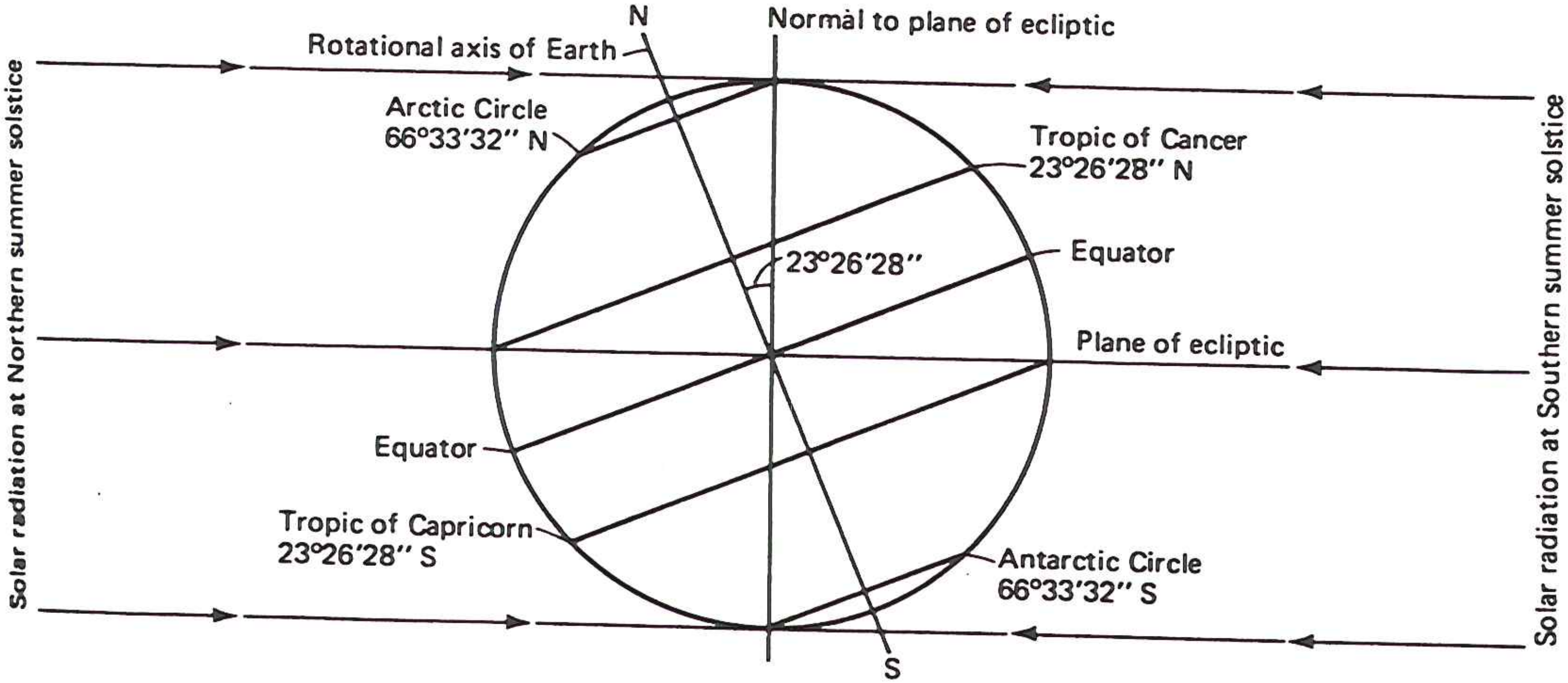


Figure 8.4. The latitudes of the tropics and the seasons are determined by the inclination of the Earth's axis from the normal to the orbital plane. As a result of this inclination, the Sun at local noon appears to move up and down during the year, with an excursion of 46°52′56″. At the vernal (spring) and fall equinoxes, the Sun stands vertical over the equator, the two hemispheres are equally lighted, and nights and days have the same duration at all latitudes except at the poles, which are grazed by the Sun's rays for 24 hours. As the Sun moves toward a tropic, the days become longer than the nights in that hemisphere, reaching a maximum of 6 months at the pole. Meanwhile, the nights become longer than the days in the opposite hemisphere, reaching a maximum of 6 months at the pole. (Emiliani, 1987, p. 216.)

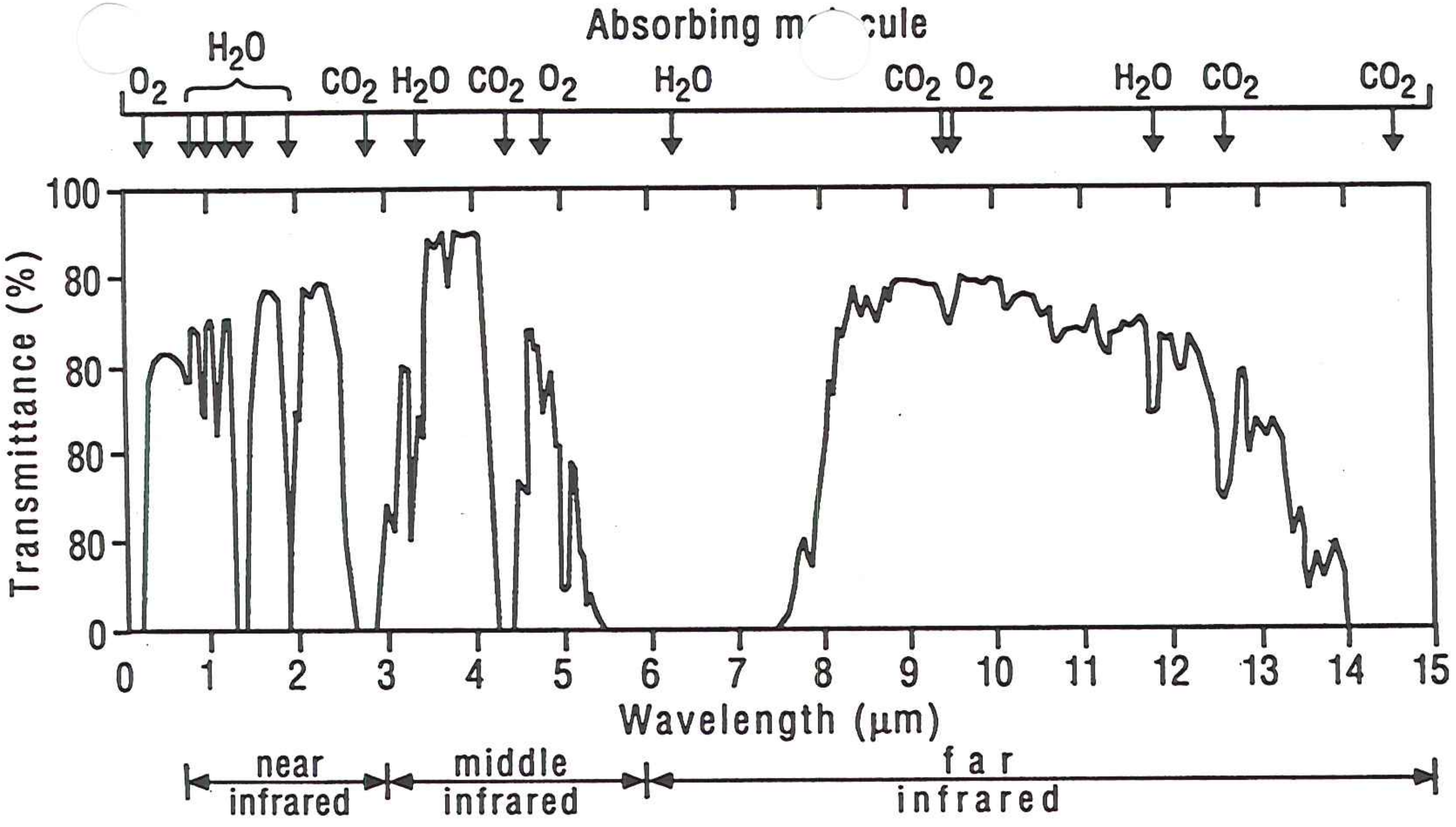
Table 6.2. Intensity of solar radiation in different wavelength bands

Type of radiation	Wavelength (μm)	Power (W/m^2)
ultraviolet	0.2–0.3	11
violet	0.3–0.4	91
blue	0.4–0.5	198
green–yellow	0.5–0.6	193
orange–red	0.6–0.7	162
red–infrared	0.7–0.8	127
infrared	0.8–0.9	100
infrared	0.9–1.0	80
infrared	1.0–1.1	66
infrared	1.1–1.2	55
infrared	1.2–1.3	45
infrared	1.3–1.4	36
infrared	1.4–1.5	30
infrared	1.5–1.6	24
infrared	1.6–1.7	20
infrared	1.7–1.8	17
infrared	1.8–1.9	14
infrared	1.9–2.0	12
infrared	>2.0	72

Table 13.2. Composition of the atmosphere

Component	Concentration	Residence time
N_2	0.78084	$4 \cdot 10^8$ y for cycling through sediments
O_2	0.20946	6,000 y for cycling through biosphere
H_2O	$(4-0.004) \cdot 10^{-2}$	—
Ar	$9.34 \cdot 10^{-3}$	largely accumulating
CO_2	$0.346 \cdot 10^{-3}$	10 y for cycling through biosphere
Ne	$1.818 \cdot 10^{-5}$	largely accumulating
He	$5.24 \cdot 10^{-6}$	$2 \cdot 10^6$ y for escape
CH_4 (methane)	$1.55 \cdot 10^{-6}$	2.6–8 y
Kr	$1.14 \cdot 10^{-6}$	largely accumulating
H	$5.5 \cdot 10^{-7}$	4–7 y
N_2O	$3.3 \cdot 10^{-7}$	5–50 y
CO	$(2-0.6) \cdot 10^{-7}$	0.5 y
Xe	$8.7 \cdot 10^{-8}$	largely accumulating
O_3 (ozone)	$(3-1) \cdot 10^{-8}$	—
CH_2O (formaldehyde)	$< 1 \cdot 10^{-8}$	—
NH_3 (ammonia)	$(20-6) \cdot 10^{-9}$	about 1 d
SO_2	$(4-1) \cdot 10^{-9}$	hours to weeks
$NO + NO_2$	10^{-9}	$<$1 mo
CH_3Cl (methyl chloride)	$5 \cdot 10^{-10}$	—
CCl_4 (carbon tetrachloride)	$(2.5-1) \cdot 10^{-10}$	—
CF_2Cl_2 (Freon 12)	$2.3 \cdot 10^{-10}$	45–68 y
H_2S	$\leqslant 2 \cdot 10^{-10}$	$\leqslant$l d
$CFCl_3$ (Freon 11)	$1.3 \cdot 10^{-10}$	45–68 y

Concentrations, by volume, of components at ground level (excluding local pollutants) and their residence times.

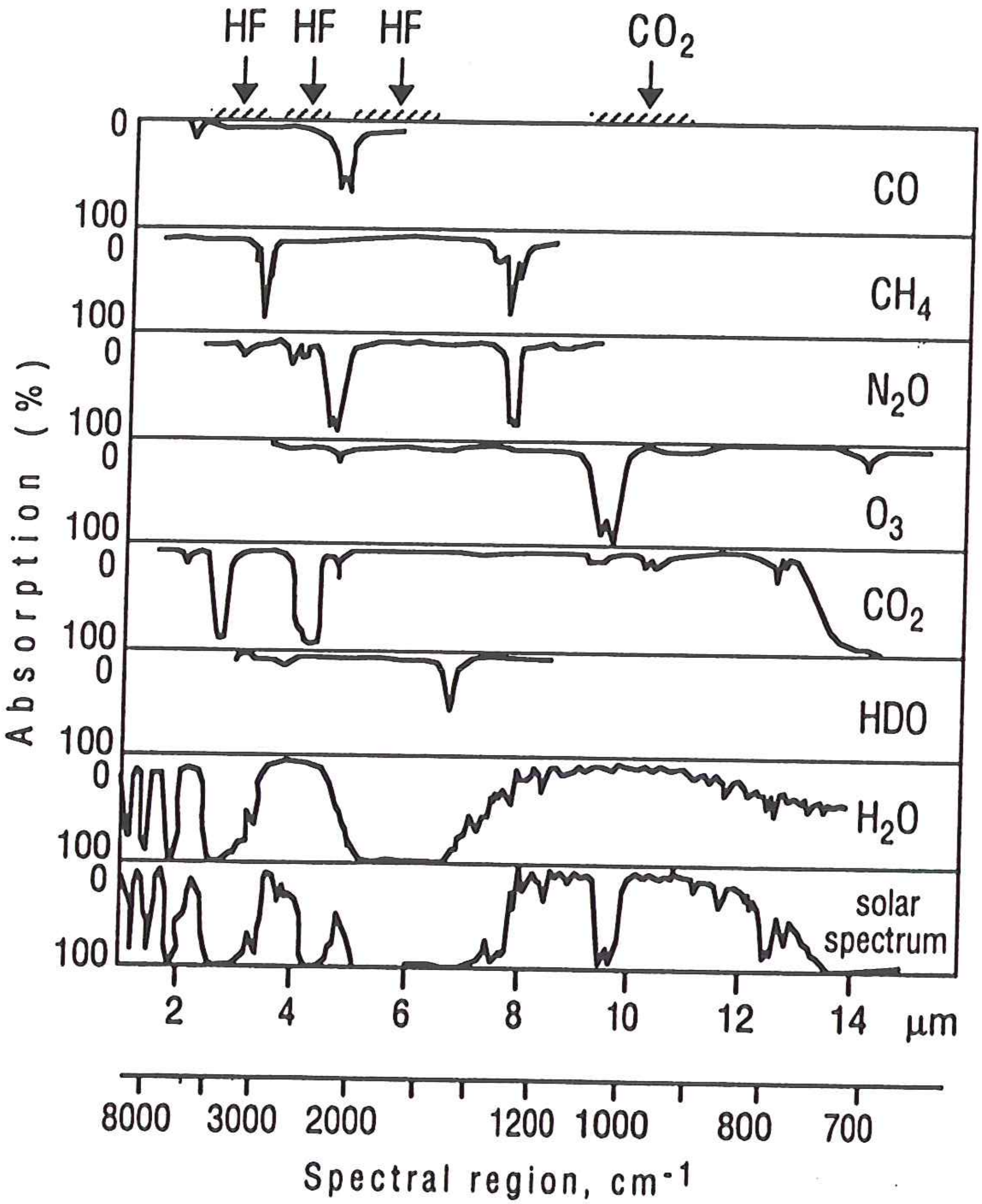


horizontal absorption at sea level

(a)

Figure 2.3 Absorption properties of the earth's atmosphere and of some of its major constitutions in the infrared region of the spectrum. (a) Horizontal absorption spectrum measured over a distance of 1.8 km at sea level. (Reproduced, with permission, from Hudson and Hudson (1975) in R. M. Measures, *Laser Remote Sensing*, Wiley-Interscience, New York, 1984, Fig. 4.6, p. 140) (b) Vertical absorption spectrum of the earth's atmosphere (bottom) and absorptions of individual compounds. (Reproduced, with permission, from M. Vergez–Delonde, Absorption des radiations infra-rouges par les gas atmosphériques, *J. Physique*, **25** (1964) 773)

(*continued overleaf*)



(b)

(continued)

integrated vertical absorption

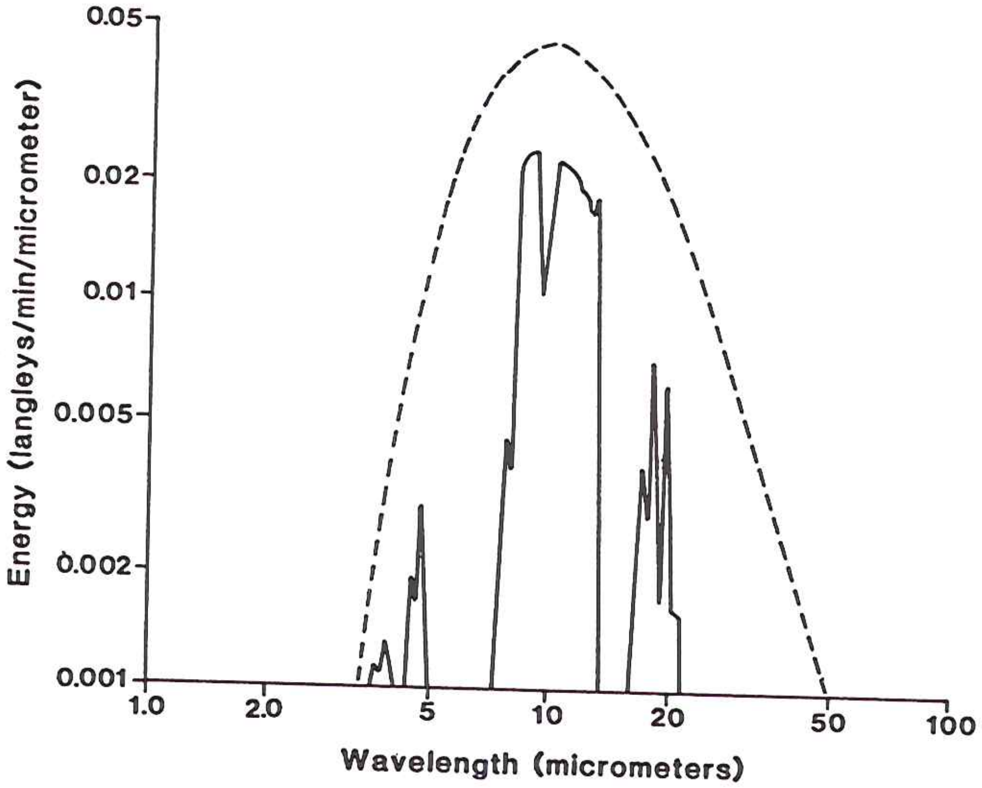
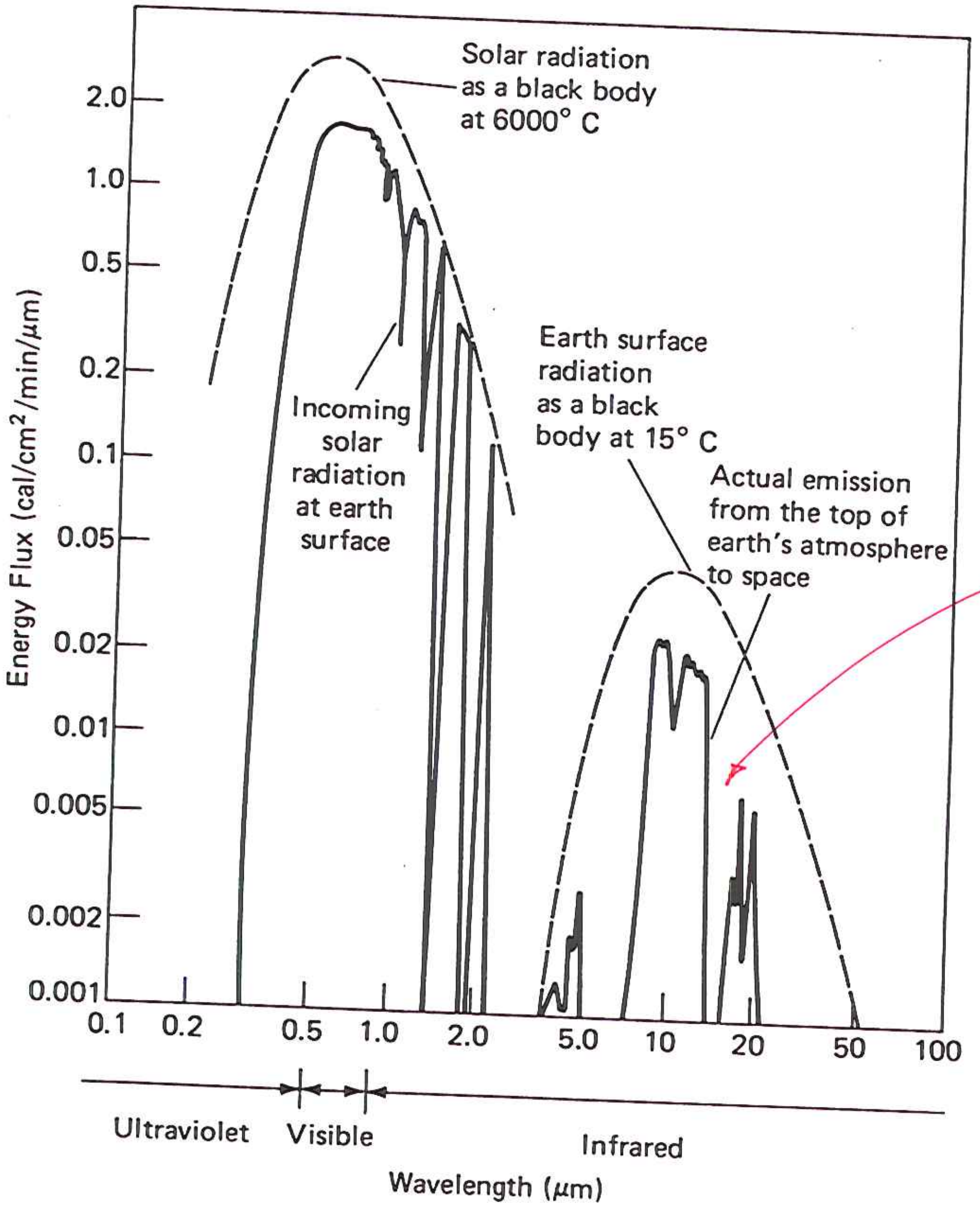


Figure 13.2. The Earth's radiation spectrum (solid line) compared with the spectrum of a blackbody radiating at 300 K (dotted line). (Adapted from Sellers, 1965, p. 21 Fig. 6.)



CO_2 band

Figure 2.5 Incoming radiation from the sun at the earth's surface and outgoing radiation from the top of the earth's atmosphere as a function of wavelength. Also shown are the energy spectra expected for black bodies at the same temperature as the sun and earth. (A black body emits the maximum radiation for its temperature.) (Adapted from W. D. Sellers, *Physical Climatology*. Copyright © 1965 by The University of Chicago Press. All rights reserved.)

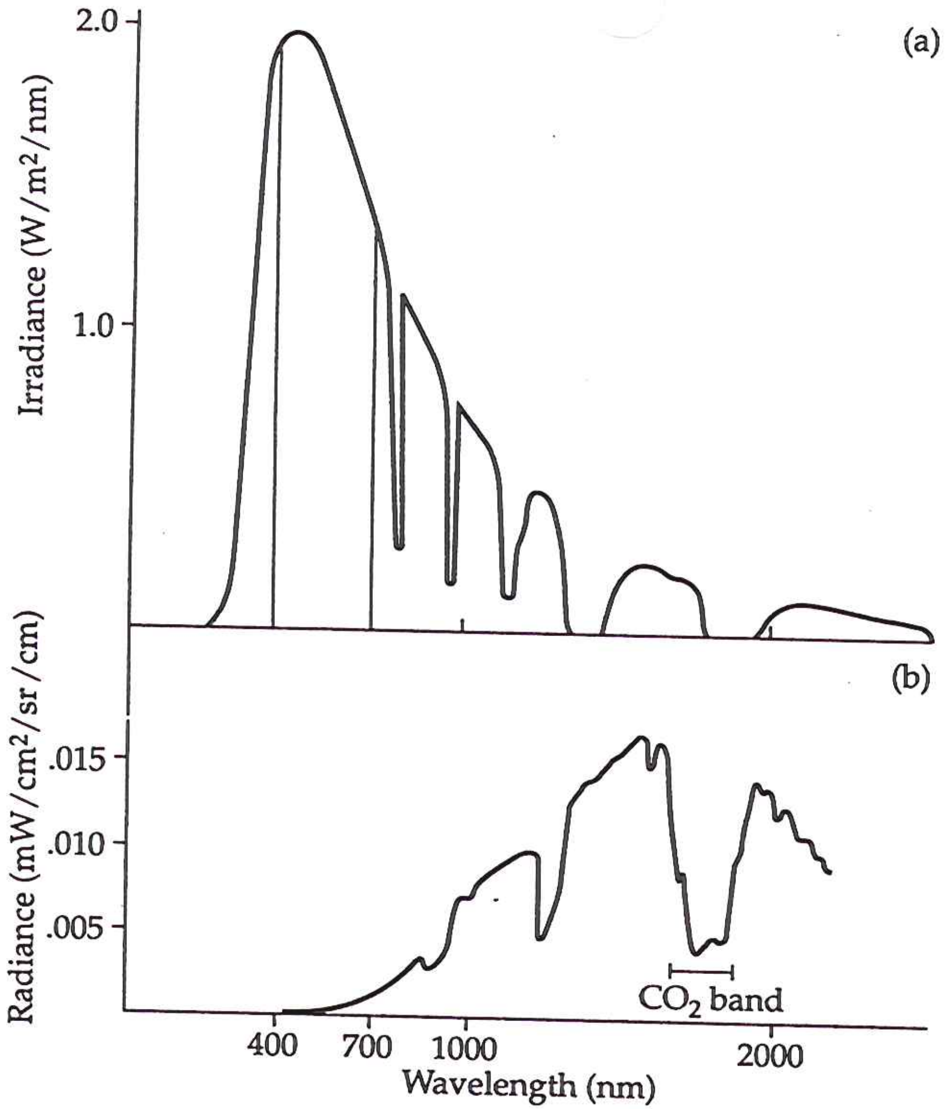


Figure 1.6 (a) The spectrum of sunlight at the earth's surface. (From Edwards and Walker 1983.) (b) The spectrum of the earth as taken from the Nimbus-4 satellite over the North African desert. (Adapted from NAS 1983.)