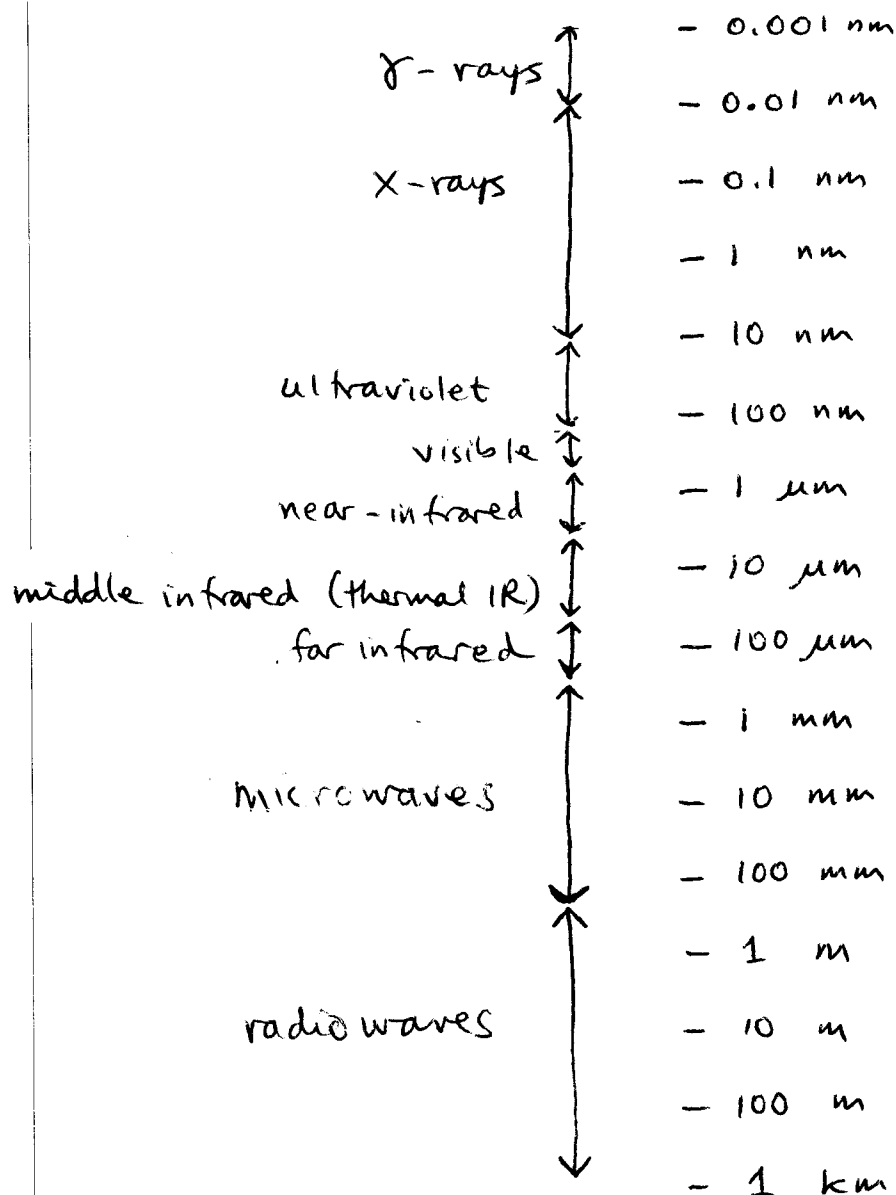


Electromagnetic radiation

is described either as waves or photons. There are no definite limits to possible wavelengths.

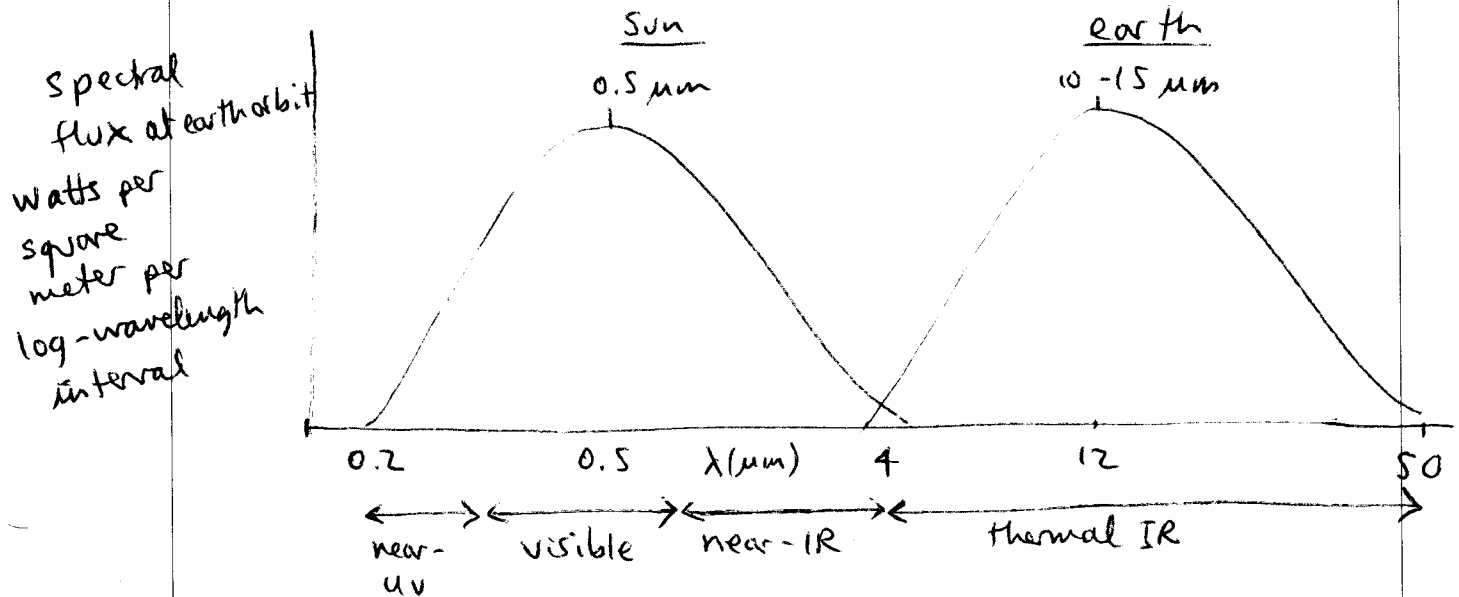


wavelength	λ	}	$\lambda \nu = c$ $\bar{\nu} = \frac{1}{\lambda}$
frequency	ν		
wavenumber	$\bar{\nu}$		
speed of light	c		

For climate, the important spectral regions are where the earth and the sun radiate significant amounts of energy:

sun: $0.2 - 4 \mu\text{m}$ "solar", "shortwave" radiation

earth: $4 - 50 \mu\text{m}$ "terrestrial", "longwave", "thermal", "infrared" radiation.



Interaction of Radiation with Matter

Working definitions:

1. Absorption: A photon is destroyed. Its energy is distributed by collisions into other forms of energy (vibrational, rotational, etc.). The frequency of subsequently *emitted* radiation is independent of the frequency of the absorbed photon. [Assuming collision-dominated; i.e. that frequency of collisional transitions is much greater than frequency of radiation-induced transitions. This is the assumption of “local thermodynamic equilibrium”.]

2. Scattering: The photon changes direction but not frequency.

3. Emission: A photon is created. The molecule emitting the photon loses internal energy (vibrational, rotational, etc.) equivalent to the energy of the photon.

4. Extinction: All processes that remove photons from a beam of light. Extinction is the sum of scattering-out and absorption.

5. A beam of light passing through a medium can be:
Depleted by *scattering-out-of-the-beam* and by *absorption*.
Augmented by *scattering-into-the-beam* and by *emission*.

Interaction of radiation with Earth’s surface and atmosphere

Absorption of solar radiation	Mostly at surface, some in atmosphere
Scattering (reflection) of solar radiation	Mostly by clouds, some at surface
Emission of infrared radiation	Surface, clouds, and atmospheric gases

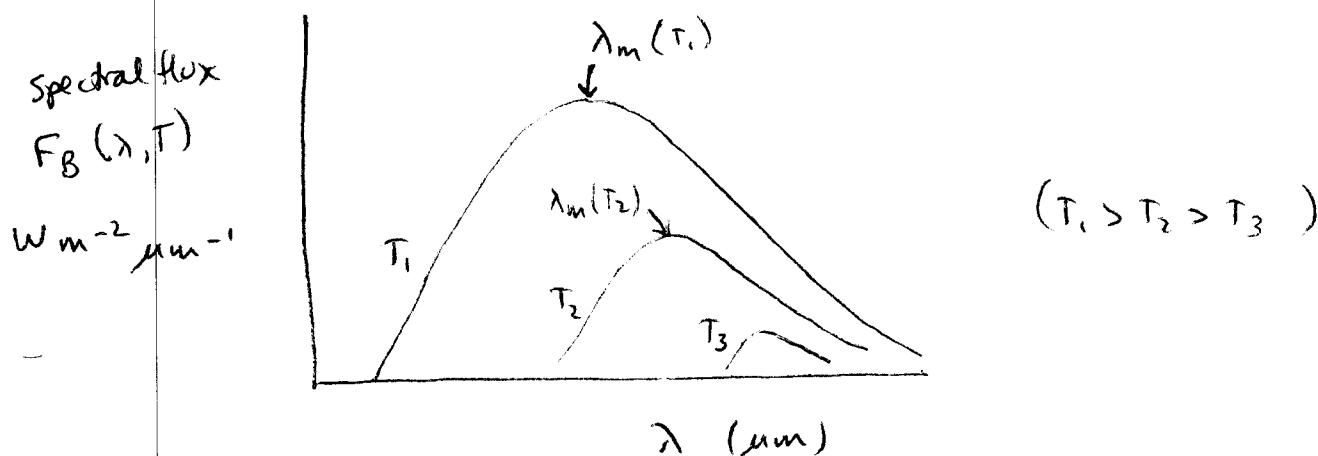
Blackbody radiation

Definition

A blackbody is an object that absorbs all radiation incident on it.

Experimental fact

The radiation emitted by a blackbody at temperature T is given by the blackbody function (Planck function) :



F_B is the spectral flux: the radiation flux emitted by a unit area of the blackbody, per unit wavelength interval.

$F_B(\lambda, T)$ depends only on the temperature of the blackbody, not on the composition of the blackbody nor on whether radiation is incident on it, and is isotropic.

The graph shows that F_B is greater at all wavelengths for a blackbody at higher temperature. It also shows that the wavelength of maximum emission occurs at shorter wavelength for higher temperature.

The functional form of F_B was discovered by Planck:

$$F_B(\lambda, T) = \frac{2\pi hc^2}{\lambda^5 (e^{\frac{hc}{\lambda kT}} - 1)} \quad \text{W m}^{-2} \mu\text{m}^{-1} \quad (1)$$

where h = Planck's constant
 k = Boltzmann's constant
 c = speed of light
 T = absolute temperature (K)

In terms of frequency:

$$F_B(\nu, T) = \frac{2\pi h \nu^3}{c^2 [e^{h\nu/kT} - 1]} \quad \text{W m}^{-2} \text{Hz}^{-1}$$

Properties of the Planck function

- ① Total radiation emitted by a blackbody is obtained by integrating (1) over wavelength:

$$F_B(T) = \int_0^\infty F_B(\lambda, T) d\lambda = \int_0^\infty \frac{2\pi hc^2}{\lambda^5 (e^{\frac{hc}{\lambda kT}} - 1)} d\lambda$$

$$= \underbrace{\frac{2\pi^5 k^4}{15 c^2 h^3}}_{\text{this is all a constant. call it } \sigma} T^4$$

this is all a constant. call it σ .

Stefan-Boltzmann law:

$$F_B = \sigma T^4$$

where $\sigma = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{K}^{-4}$.

② Wavelength of maximum emission

Given by "Wien's displacement law" (The maximum is displaced as the temperature changes).

$$\frac{dF_B(\lambda, T)}{d\lambda} = 0$$

$$\frac{d}{d\lambda} \left[\frac{2hc^2}{\lambda^5 (e^{hc/\lambda kT} - 1)} \right] = 0 \quad \text{at } \lambda = \lambda_{\max}$$

$$\text{so } \lambda_{\max} = \frac{2897 \mu\text{mK}}{T}$$

so an observation of $\lambda_{\max}$ (for a blackbody) tells T .

eg. estimate the temperature of the sun and the earth by observing $\lambda_{\max}$.

	$\lambda_{\max}$	$\Rightarrow$	T
sun	0.475 μm		6100 K
earth	11 μm		255 K.

The Solar Spectrum

The distribution of electromagnetic radiation emitted by the sun approximates black-body radiation for a temperature of about 6000 K as shown in Fig. 5.6. The similarity between the solar and black-body spectra provides the basis for estimates of the temperature of the visible surface layer of the sun. Because the sun does not radiate as a perfect black body, different comparisons are possible with different results. The Stefan-Boltzmann law may be used in combination with the solar constant, as in Problem 5, to calculate a temperature of 5733 K. Or the wavelength of maximum intensity, $0.4750 \mu\text{m}$, combined with Wien's displacement law [Eq. (5.20)] may be used to calculate a temperature of 6100 K. Although it is not pos-

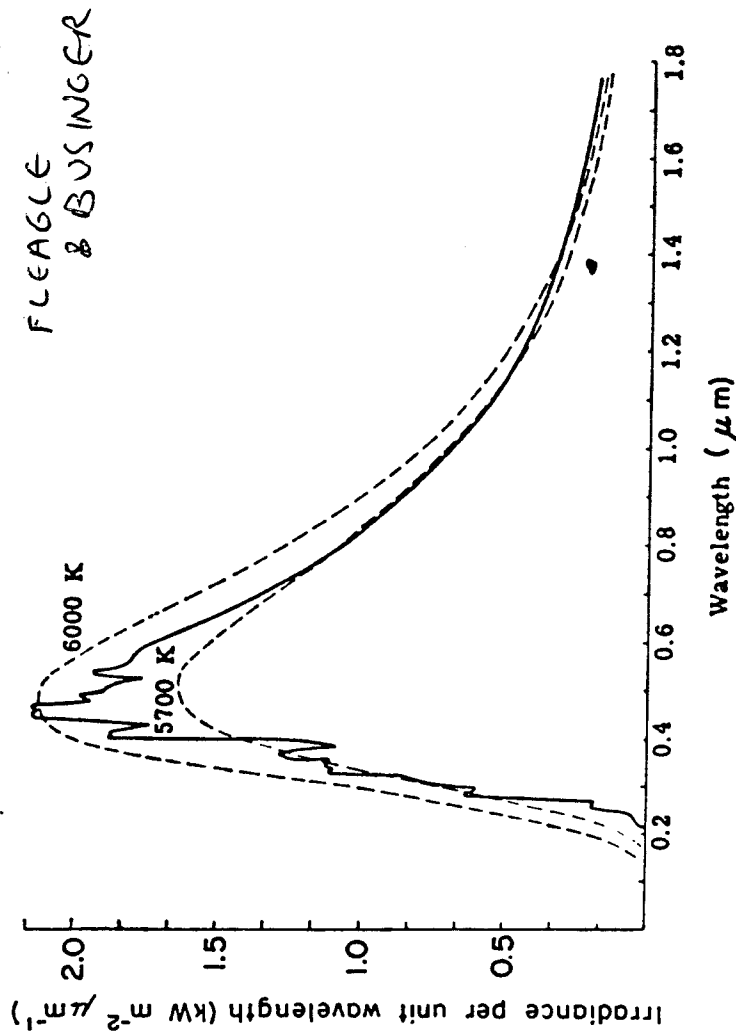


FIG. 5.6. Observed solar spectral irradiances corrected to mean solar distance (solid line) and corresponding black-body spectral irradiances for temperatures of 6000 and 5700 K (dashed lines) (after F. S. Johnson, *J. Meteorol.* 11, 431, 1954).

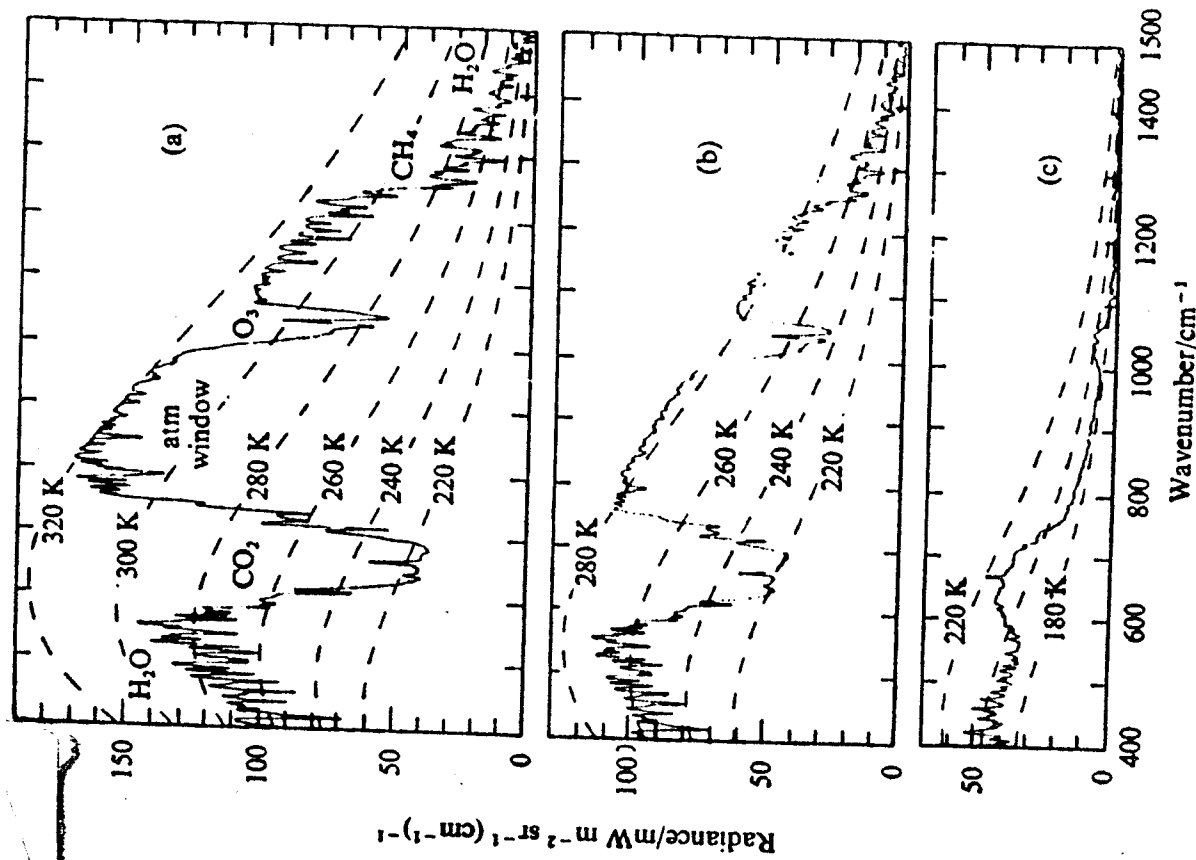
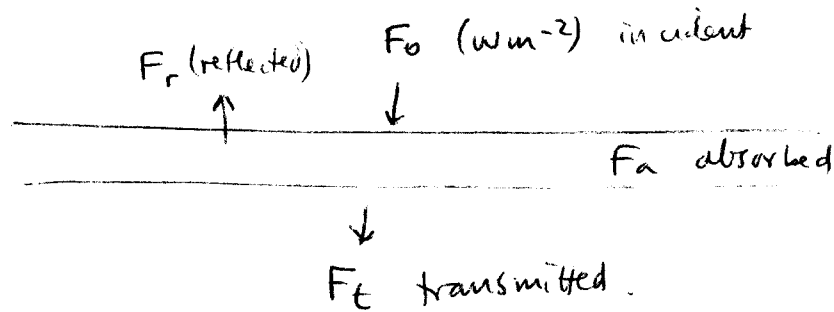


Fig. 12.3. Thermal emission from the earth plus atmosphere emitted vertically upwards and measured by the infrared interferometer spectrometer on Nimbus 4, (a) over Sahara, (b) over Mediterranean, (c) over Antarctic. The radiances of black bodies at various temperatures are superimposed. (From Hanel *et al.*, 1971)

Radiative properties of nonblack materials

Consider radiation F_0 incident on one side of a horizontally homogeneous slab of some material (like a stratus cloud, or the ozone layer, or a snowpack)



In steady state, F_0 , F_r , F_a and F_t are constant ^{in time}, and

$$F_0 = F_r + F_a + F_t$$

Define

transmittance (transmissivity)

$$t = F_t / F_0$$

reflectance (reflectivity, albedo)

$$r = F_r / F_0$$

absorptance (absorptivity)

$$a = F_a / F_0$$

so
$$F_0 = r F_0 + a F_0 + t F_0$$

$$1 = r + a + t.$$

These ratios can vary with wavelength and are all between 0 & 1.

$$0 \leq r(\lambda) \leq 1$$

$$0 \leq a(\lambda) \leq 1$$

$$0 \leq t(\lambda) \leq 1$$

$$r(\lambda) + a(\lambda) + t(\lambda) = 1.$$

For opaque surface: $r(\lambda) + a(\lambda) = 1$, $t = 0$

For blackbody: $a(\lambda) = 1$, $r = t = 0$

Kirchhoff's law

describes emission from nonblack materials.

Let $F(\lambda)$ ^(Wm⁻²) be the emission rate from an object

(which may be nonblack) at wavelength λ at temperature T .

Let $a(\lambda)$ be the absorptance of the object.

Let $F_B(\lambda)$ be the emission rate from a blackbody at the same temperature at wavelength λ .

Define emittance (emissivity) $\epsilon(\lambda) \equiv \frac{F(\lambda)}{F_B(\lambda)}$

Kirchhoff's law: (valid for thermal equilibrium⁺
or in a collision-dominated system)

$$\epsilon(\lambda) = a(\lambda)$$

$$\text{So } F(\lambda) = a(\lambda) F_B(\lambda)$$

Since $a \leq 1$, nothing in thermal equilibrium
can emit more than a blackbody.

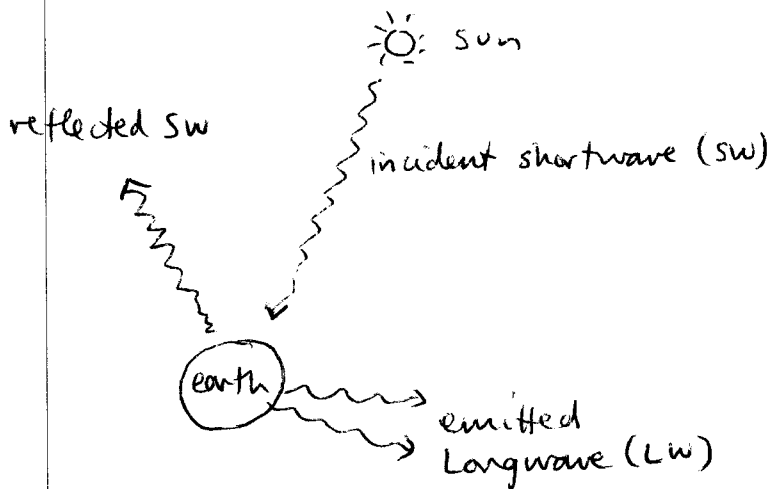
For a blackbody, $a = \epsilon = 1$

For an opaque surface $a(\lambda) + r(\lambda) = 1$

$$\text{so } \epsilon(\lambda) + r(\lambda) = 1$$

Planetary radiation balance

(45)



$$\text{Incident SW} = S_0 \pi r_e^2 \quad \text{Watts}$$

$$\text{where } S_0 = 1370 \text{ Wm}^{-2}$$

and $r_e = \text{radius of earth}$.

Reflected SW

Define albedo

(symbol α)

$$\alpha = \frac{\text{SW} \uparrow}{\text{SW} \downarrow}$$

Albedo can be measured at any level: at surface α_s , at top of cloud α_c , at top of atmosphere α_p . For planetary radiation balance we need the albedo at top of atmosphere, also called "planetary albedo" α_p .

Typical albedos for earth surfaces:

forest	$\alpha_s \approx 0.1$
snow	$\alpha_s \approx 0.8$
dry soil	$\alpha_s \approx 0.2$
cloud	$\alpha_c \approx 0.4$ average
	0.1-0.9 range

i.e. SW reflectivity of earth surfaces are quite variable

The average planetary albedo for the earth is measured to be

$$\alpha_p = 0.3.$$

This can be explained as an average of cloudy areas and cloud-free areas.

cloud-free planetary albedo ≈ 0.15

cloudy-sky planetary albedo ≈ 0.4

average cloud cover of earth ≈ 0.6

$$\text{so } \alpha_p = (0.6)(0.4) + (1 - 0.6)(0.15) = 0.3$$

$$\text{Reflected solar radiation} = S_0 \pi r_e^2 \alpha_p$$

$$\text{Absorbed solar radiation} = \text{incident SW} - \text{reflected SW}$$

$$= S_0 \pi r_e^2 (1 - \alpha_p) \quad (\text{Hartmann } \text{eq. 2.7})$$

Emitted longwave radiation

Define effective temperature of a planet T_e :

The temperature a planet would have, in radiation balance, if it emits as a blackbody in the longwave (it is not black in the shortwave; has reflectivity α_p).

So the planet emits $\sigma T_e^4 \text{ Wm}^{-2}$ over an area $4\pi r_e^2$

so emits $4\pi r_e^2 \sigma T_e^4$ (Hartmann 2.8)

Radiation balance:

Absorbed SW = Emitted LW

$$S_0 \pi r_e^2 (1 - \alpha_p) = 4 \pi r_e^2 \sigma T_e^4$$

so $\boxed{\frac{S_0}{4} (1 - \alpha_p) = \sigma T_e^4}$ (Hartmann eq. 2.9)

This is the equation for planetary radiation balance and is the definition of T_e

For the earth:

$$\alpha_p = 0.3$$

$$S_0 = 1370 \text{ Wm}^{-2}$$

$$\frac{S_0}{4} (1 - \alpha_p) = 240 \text{ Wm}^{-2}$$

$$\sigma T_e^4 = 240 \text{ Wm}^{-2}$$

$$T_e = 255 \text{ K.}$$

LW Emissivity of earth surfaces

In contrast to the situation in SW, the LW emissivity of earth surfaces are all very similar, close to 1.0:

	<u>ϵ (LW)</u>
snow	0.99
cloud	0.99
ocean	0.96
desert sand	0.9

so it is reasonable to assume earth surfaces and clouds are black in the longwave (but not in the shortwave!)

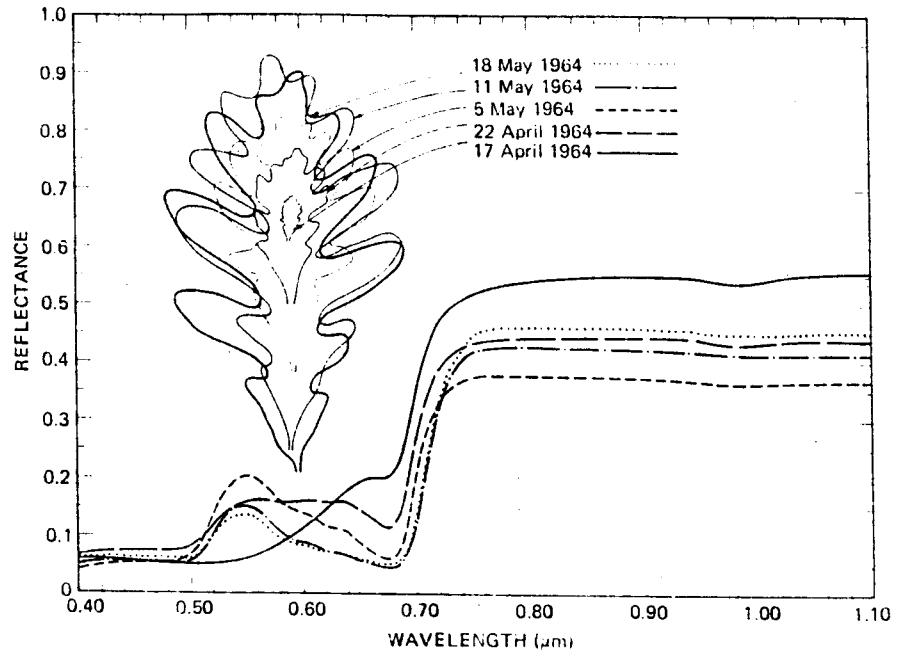
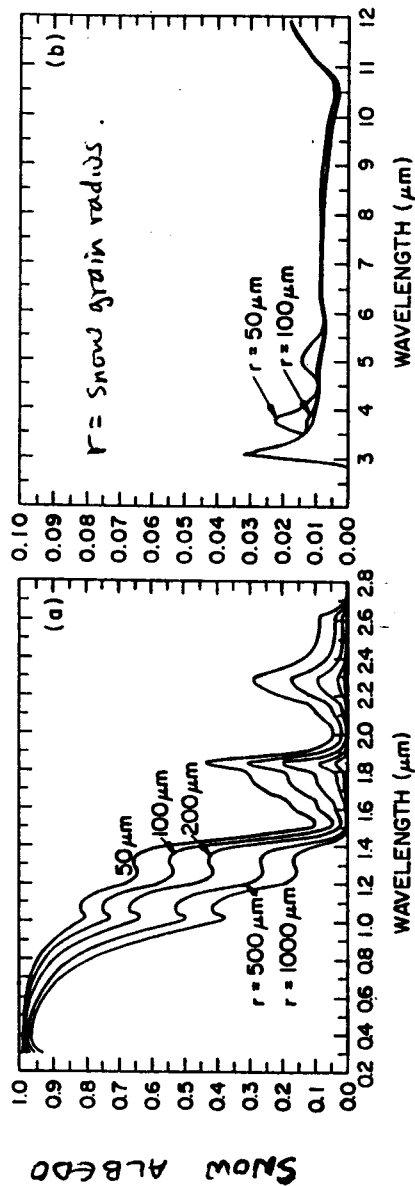


Figure 8.29. Spectral reflectance as a function of wavelength for a white oak (*Quercus alba*) leaf for different times of year.

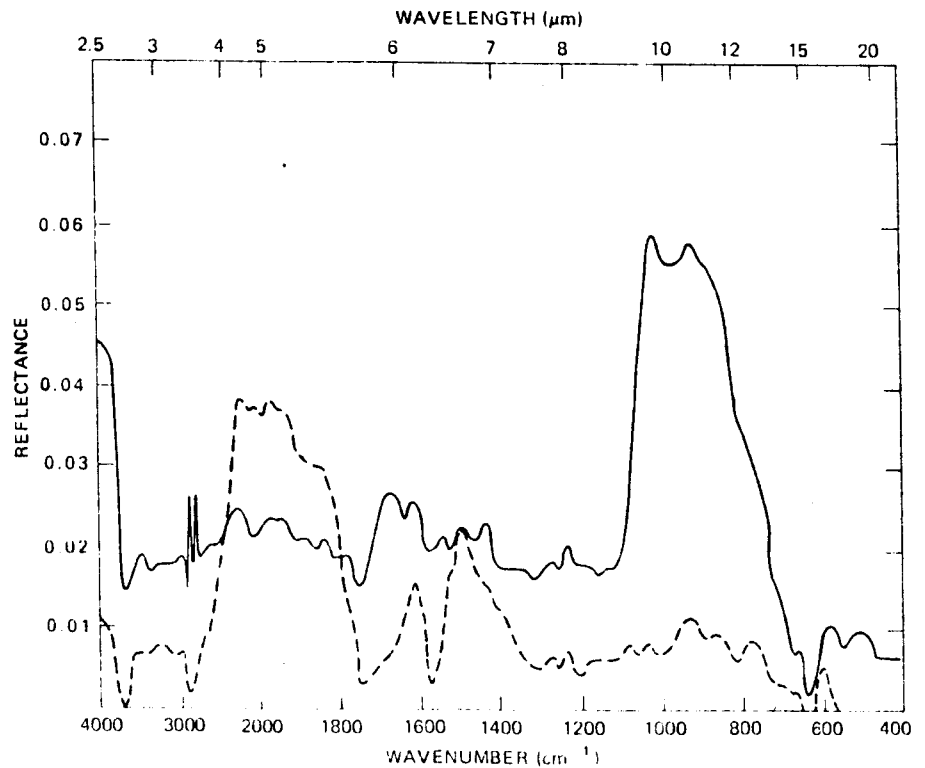
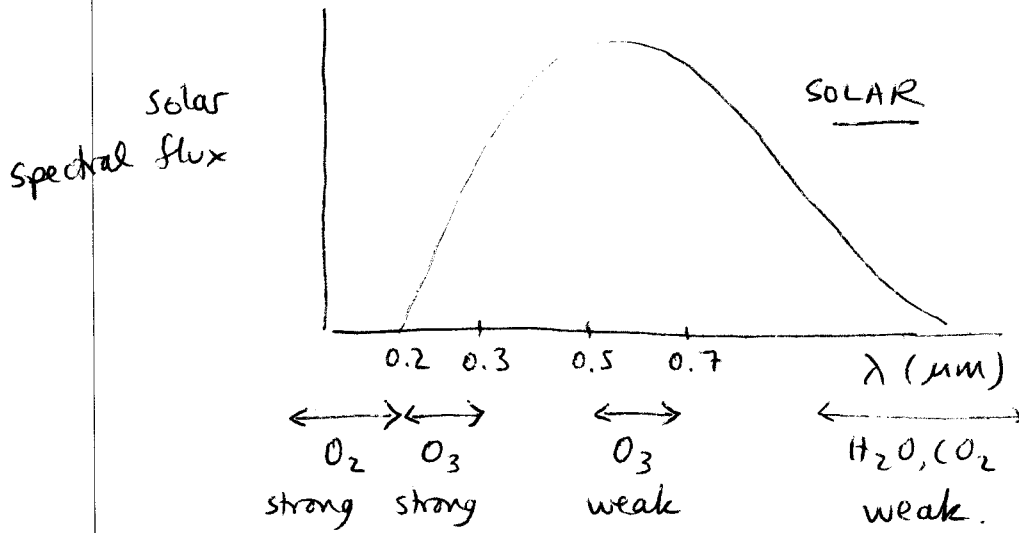
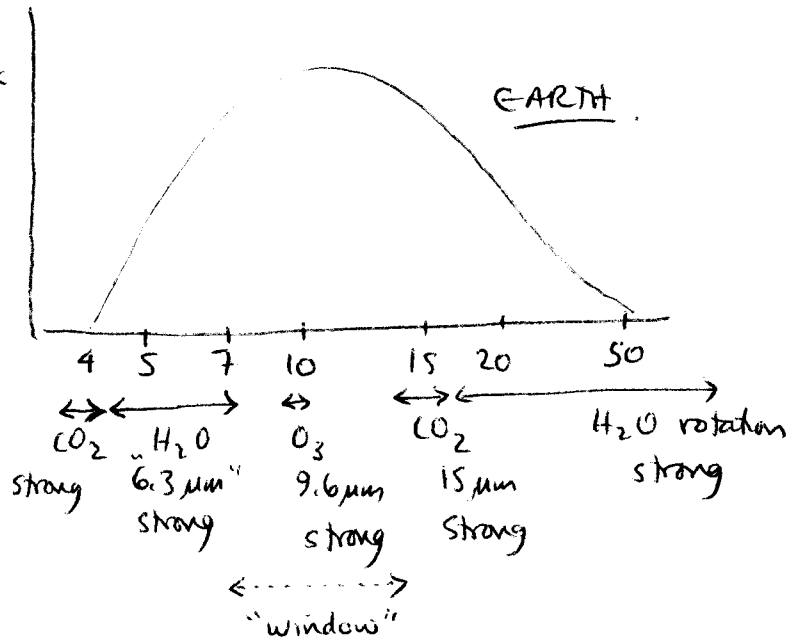


Figure 8.33. Infrared spectral reflectance as a function of wavenumber for (---) black spruce (*Picea mariana*) and (—) sugar maple (*Acer saccharum*) leaves.

Absorption of SW and LW radiation by gases on earth.



terrestrial radiation flux emitted by earth's surface



summary

UV: strong absorption by O_2 , O_3 (electronic). No solar radiation reaches the surface for $\lambda < 0.3 \mu\text{m}$

visible, near-IR: weak absorption by O_3 , H_2O , CO_2 .
Most solar radiation reaches the surface.

thermal IR: strong absorption by H_2O , O_3 , CO_2 . Atmosphere is opaque to thermal IR except at 8-13 μm . Radiation emitted by the surface escapes to space, only if it is in the window 8-13 μm . Otherwise it is absorbed by atmosphere.

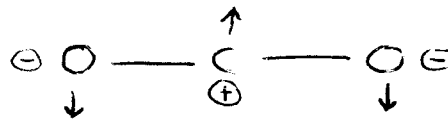
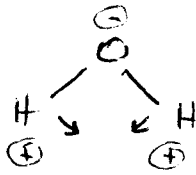
Atmospheres of the Inner Planets

	Mercury	Venus	Earth	Mars
Atmospheric pressure at surface	~0	90 atm.	1 atm = 1013 millibars (mb)	0.007-0.010 atm. (7-10 mb) Varies with season
Surface air Temperature (°C)	-170 night +330 day	480	Average 15 Range -89 to +57	Average -57 Range -130 to +30
Major gases	-	CO ₂ 97%	N ₂ 78% O ₂ 21%	CO ₂ 95%
Minor gases	-	N ₂ 2%	Ar 1%	Ar 1.6% N ₂ 2%
Gases that absorb infrared radiation (“greenhouse gases”)	-	CO ₂	H ₂ O water CO ₂ carbon dioxide O ₃ ozone CH ₄ methane N ₂ O nitrous oxide	CO ₂ H ₂ O
Notes		Covered with clouds of H ₂ SO ₄	Radiatively important gases are not the major gases.	Pressure varies because CO ₂ condenses out on polar caps.

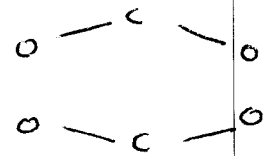
why are the major gases on earth not important for radiation?

Gases absorb infrared radiation by vibration if they have a dipole moment which can change while they vibrate.

Dipole moment: center of mass of the protons is at a different location than the center of mass of the electrons.



H_2O has a dipole moment. CO_2 has no dipole moment at rest but does when it is vibrating.



amounts of radiative gases in earth's atmosphere

CO_2 is uniformly mixed ~350 ppm or 0.035 %

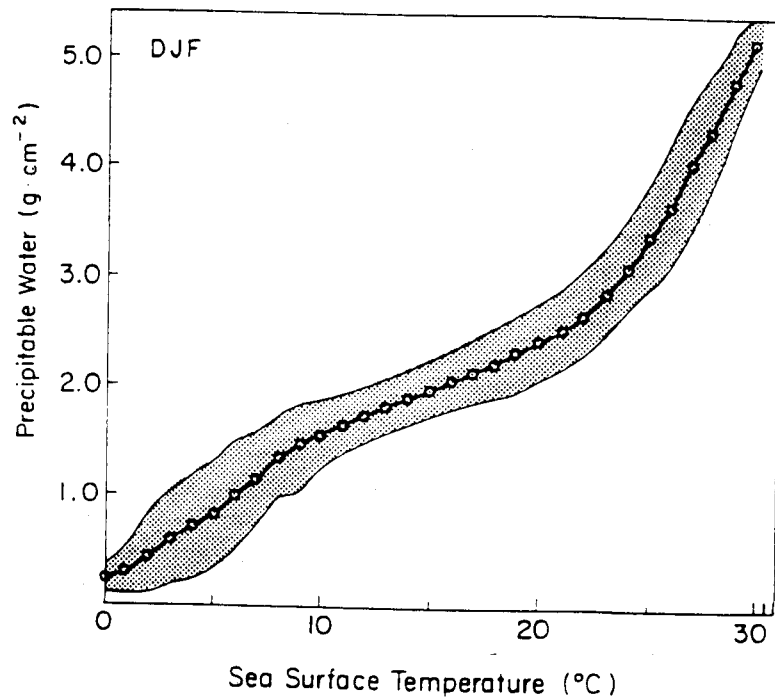
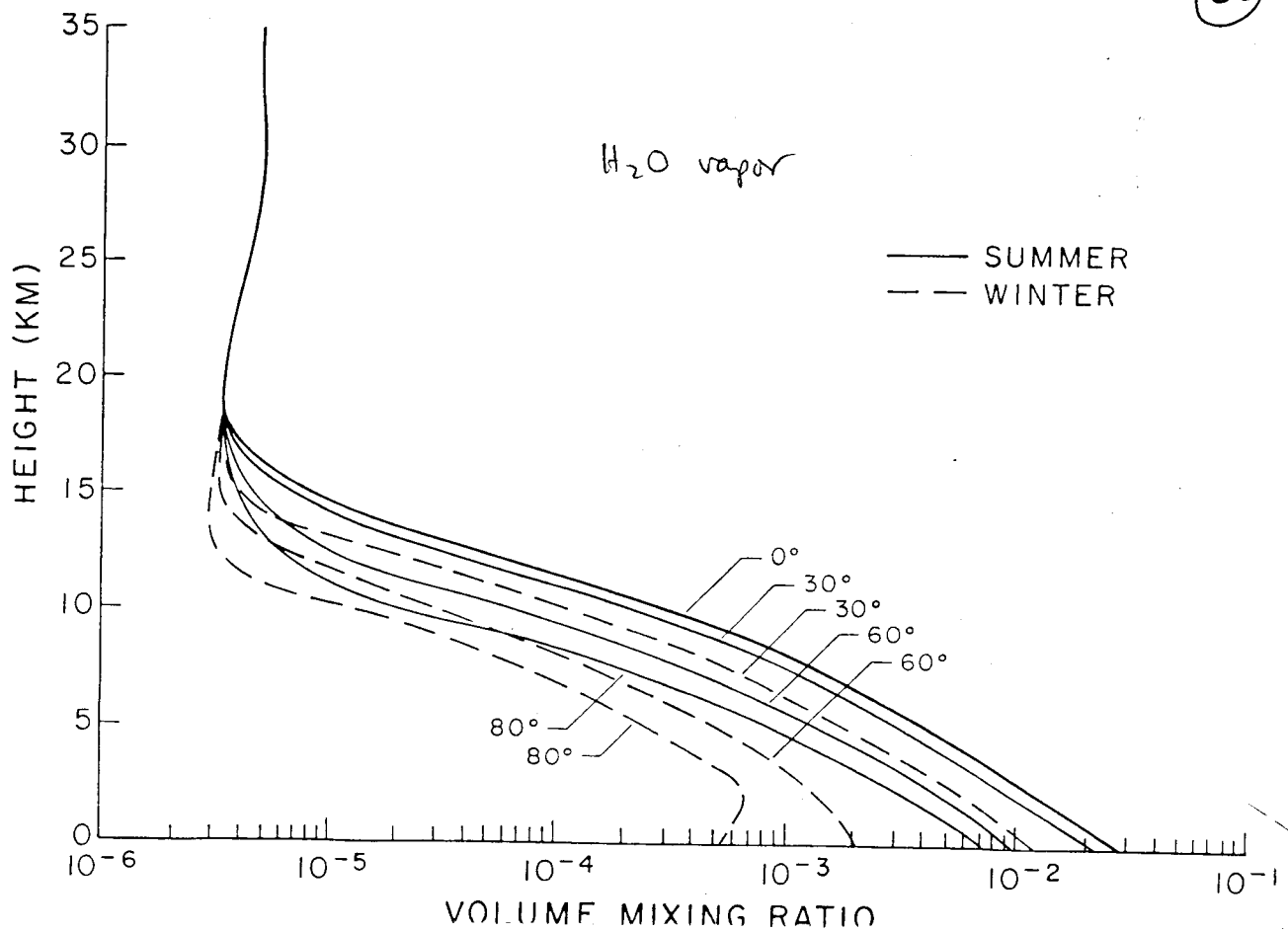
H_2O not uniformly mixed. column amount 0-5 cm "precipitable water"

O_3 not uniformly mixed. column amount 3 mm STP

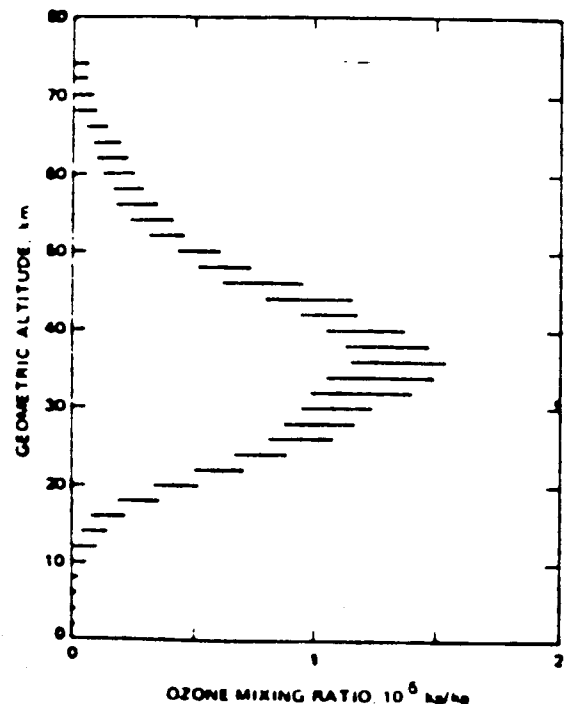
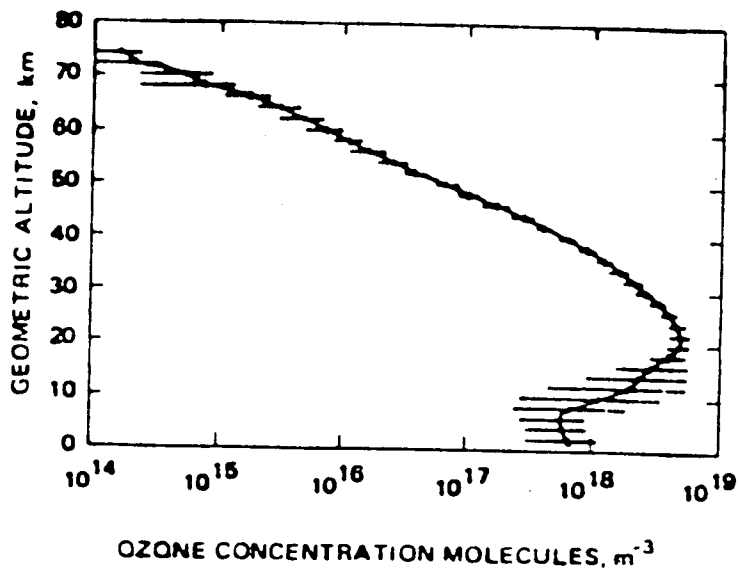
other atmospheric constituents (besides gases) important for climate

clouds

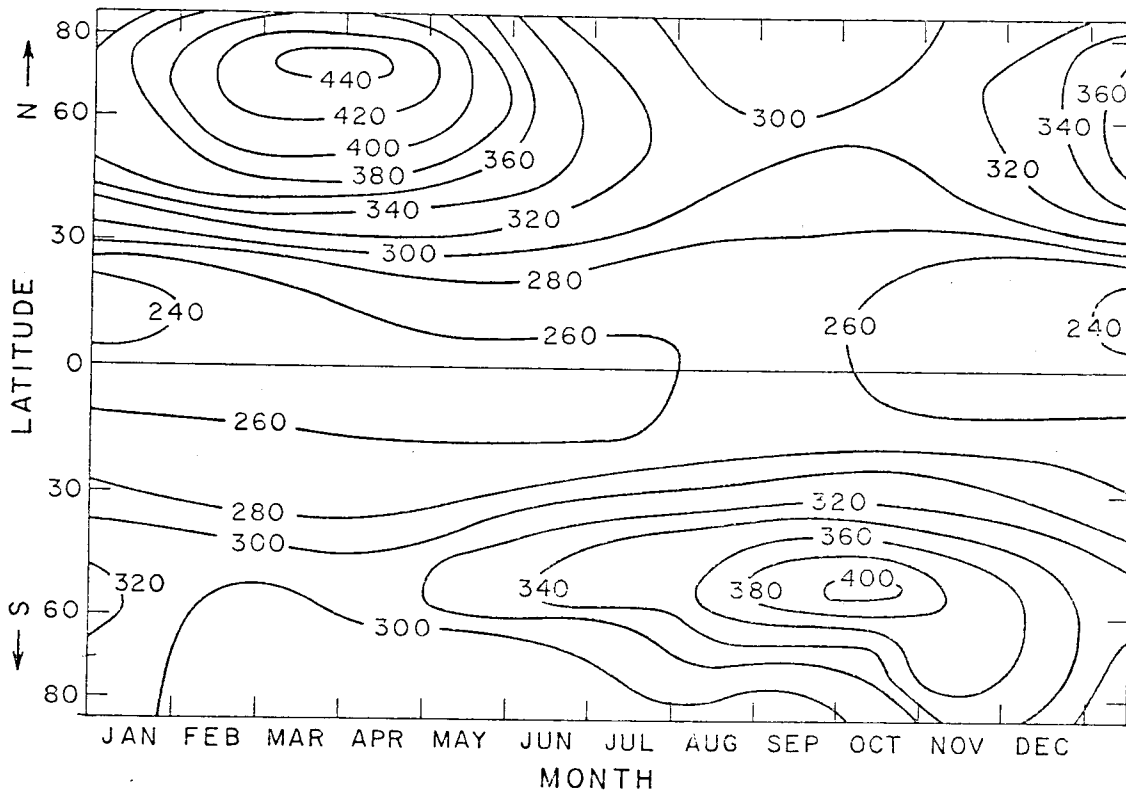
aerosols: soil dust
volcanic ash
carbon soot
sulfates.



GRAEME L. STEPHENS



$10^{-5} kg/kg$



Latitude-Season Cross Section of Total Ozone (1957-1975)

"Dobson Units": milli-centimeters STP. Divide by 100 to get mm STP.