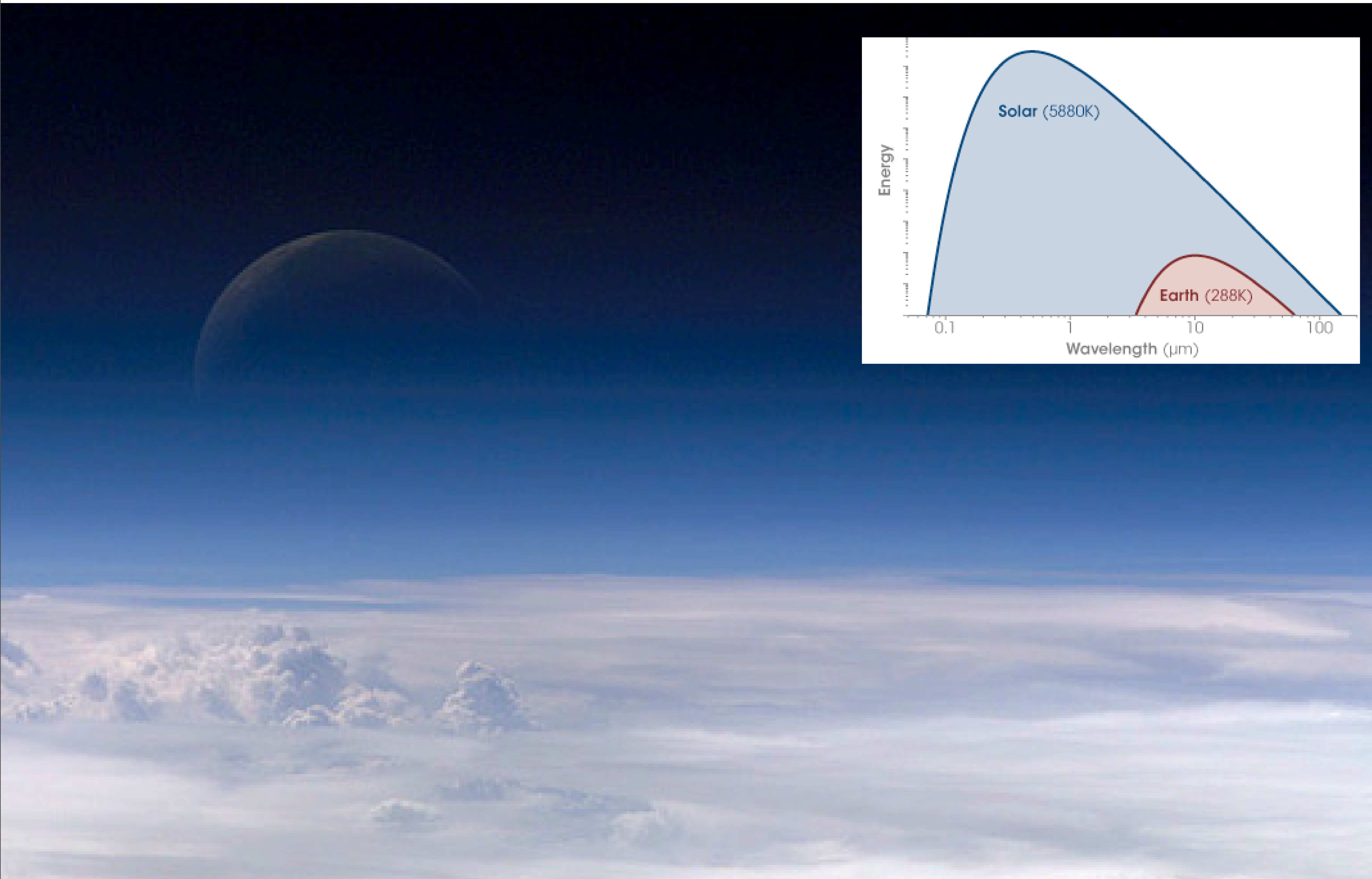
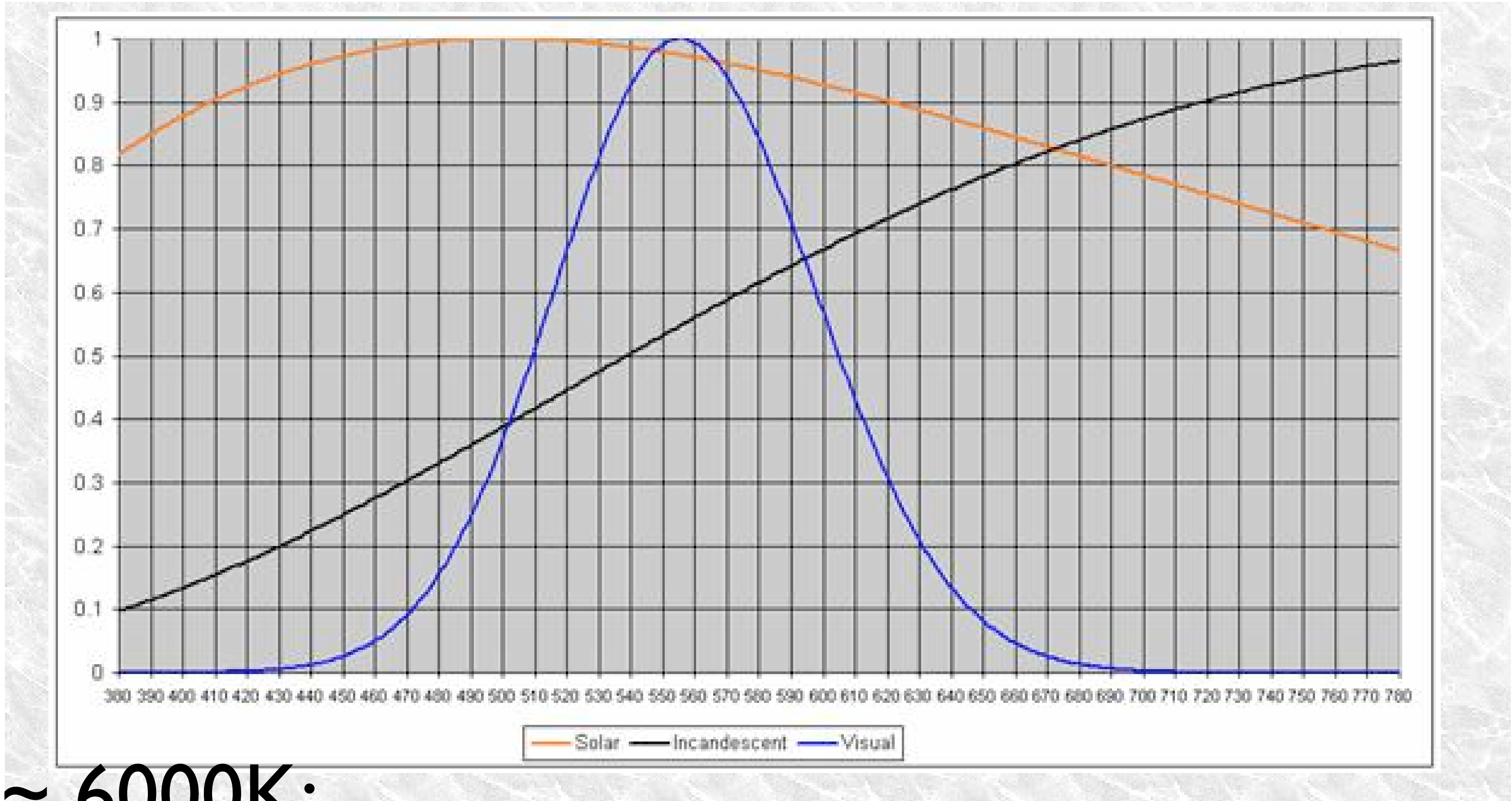
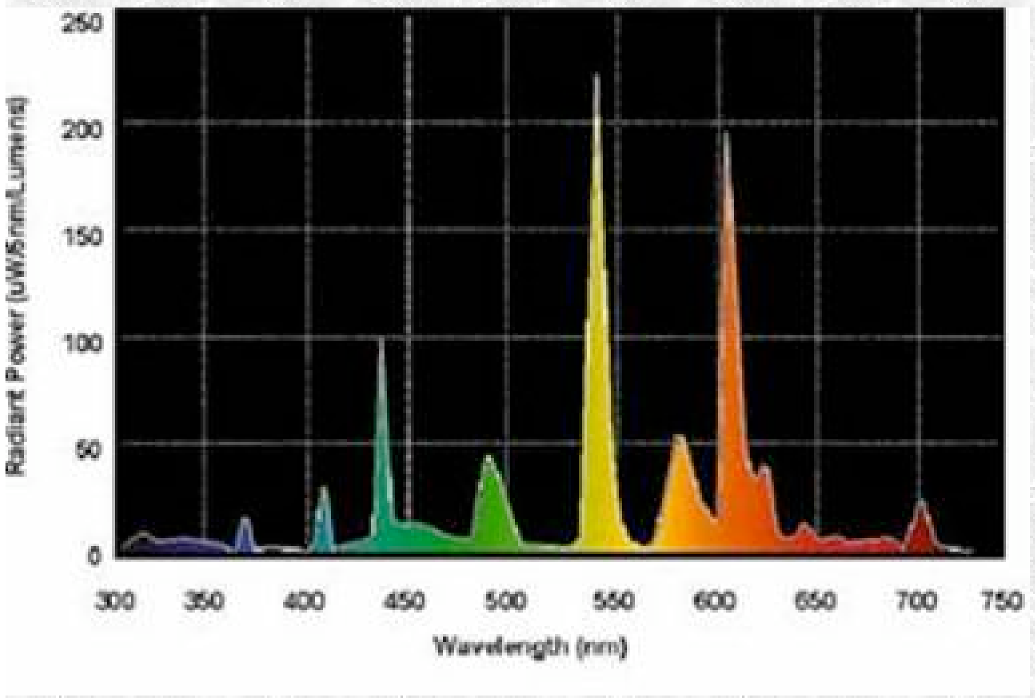


ISS013E54329 Gateway to Astronaut Photography of Earth



T_{sun} ~ 6000K;

$T_{incandescent}$ ~ 3300K => max. wavelength ~ 0.88 micron



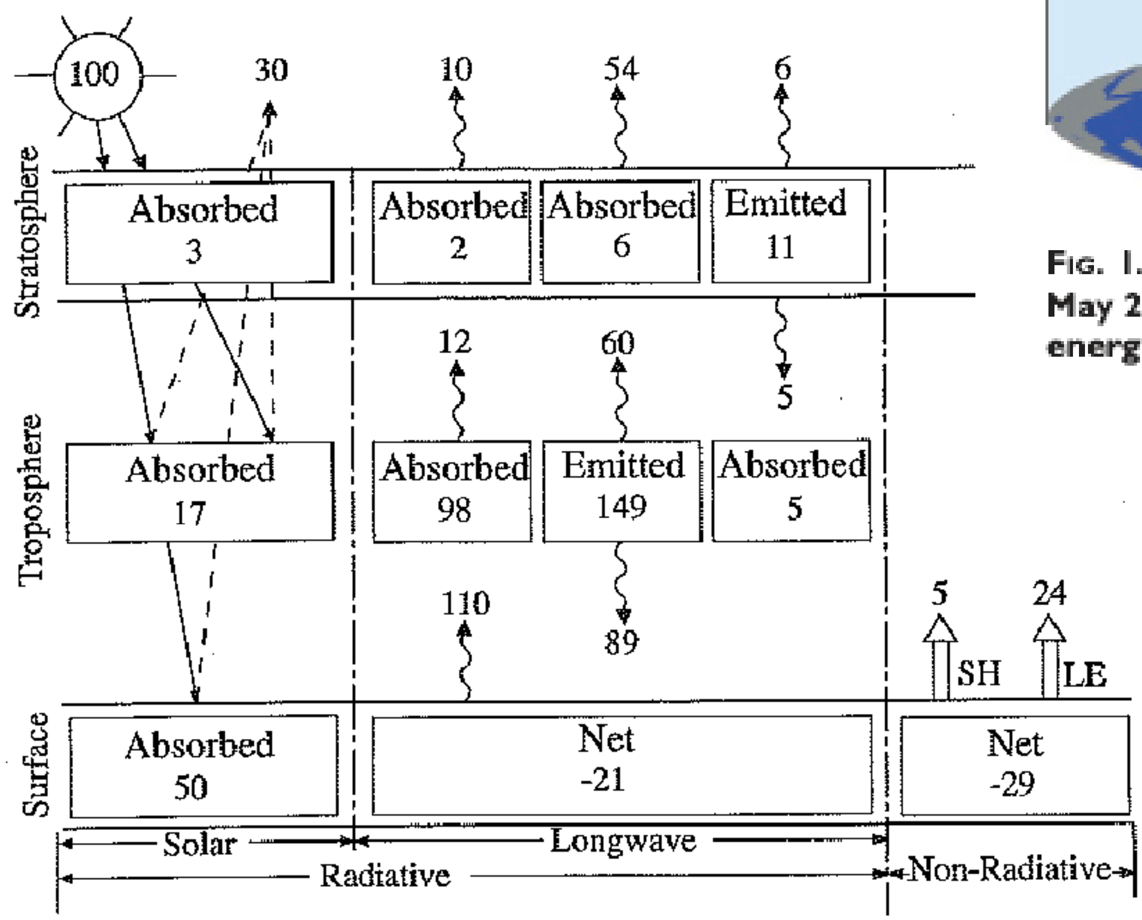


Fig. 2.4 Radiative and nonradiative energy flow diagram for Earth and its atmosphere. Units are percentages of the global-mean insolation (100 units = 342 W m^{-2}).

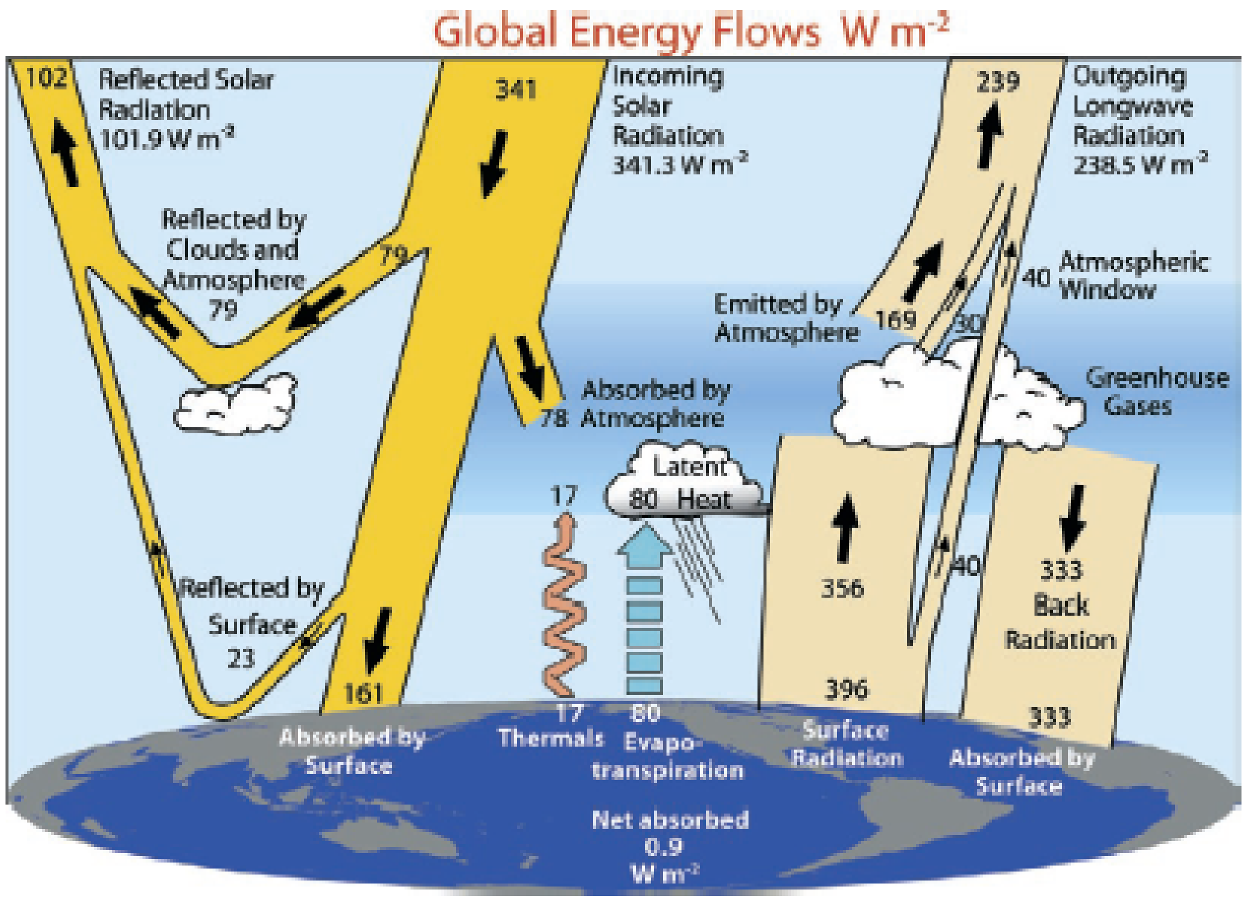


Fig. 1. The global annual mean Earth's energy budget for the Mar 2000 to May 2004 period (W m^{-2}). The broad arrows indicate the schematic flow of energy in proportion to their importance.

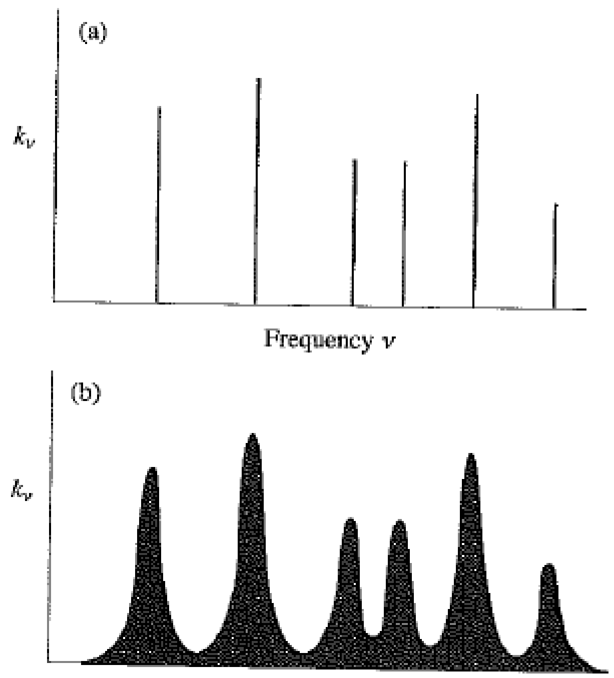
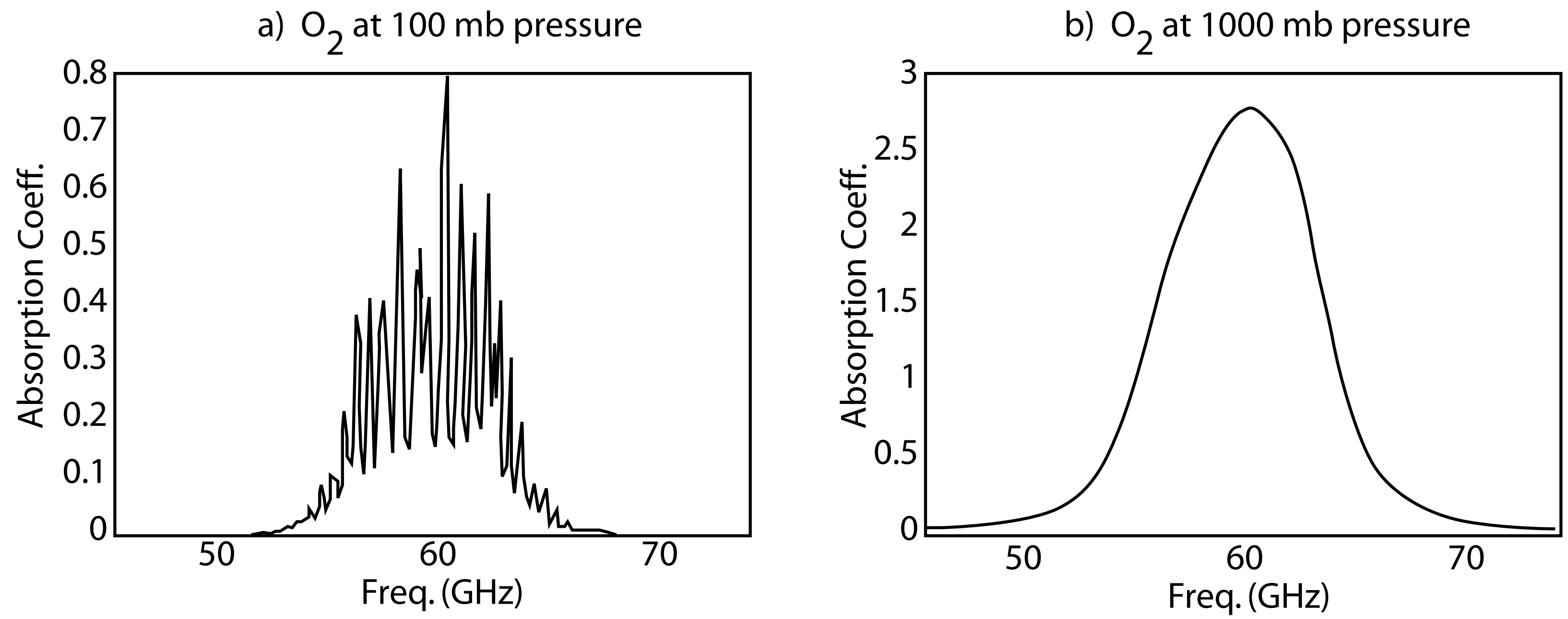


Fig. 3.5 Hypothetical line spectrum (a) before broadening, (b) after broadening.



absorption cross-section

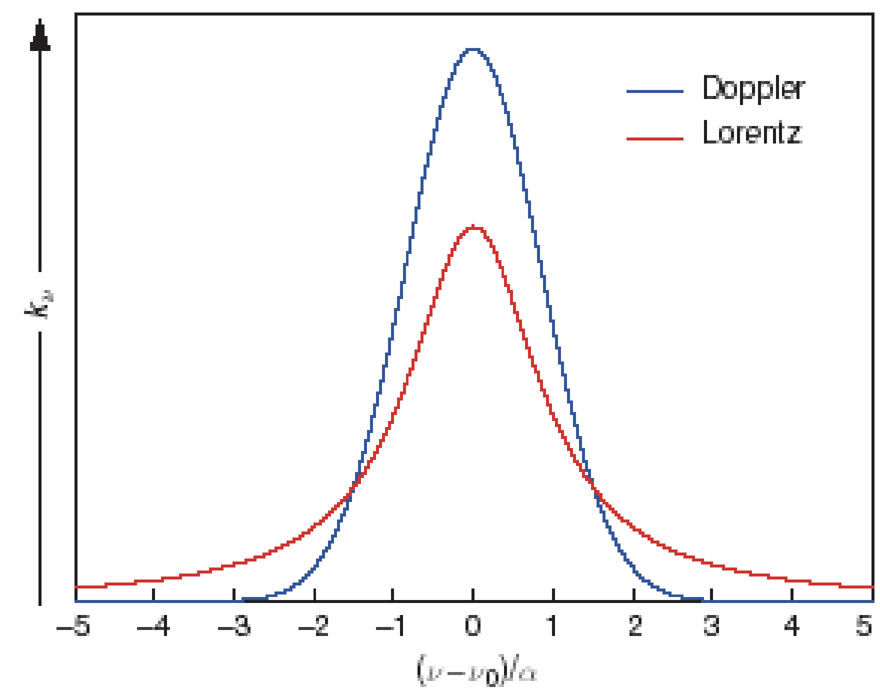
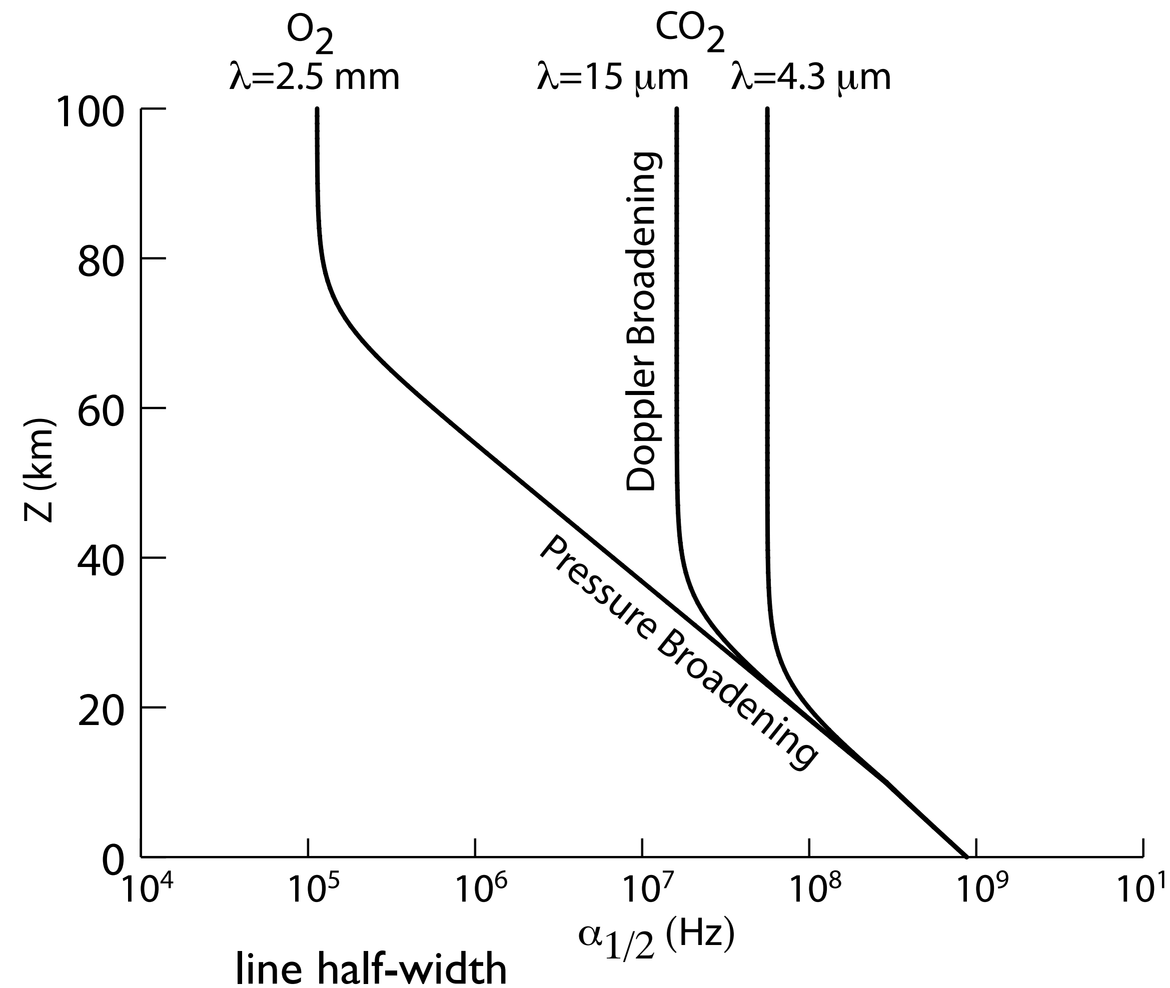


Fig. 4.21 Contrasting absorption line shapes associated with Doppler broadening and pressure broadening. Areas under the two profiles, indicative of the line intensity S, are the same. [Courtesy of Qiang Fu.]



radiative equilibrium in a 2-layer atmosphere

3.8 Heuristic Model of Radiative Equilibrium

A layer of atmosphere that is almost opaque for longwave radiation can be crudely approximated as a blackbody that absorbs all terrestrial radiation incident on it and emits like a blackbody at its temperature. For an atmosphere with a large infrared optical depth, the radiative transfer process can be represented with a series of blackbodies arranged in vertical layers. Two layers centered at 0.5- and 2.0-km altitudes provide a simple approximation for Earth's atmosphere.[1] If we assume that the atmospheric layers are transparent to solar radiation, we have the schematic energy flow diagram shown in Fig. 3.10.

We can solve for all of the unknown temperatures by using the energy balance at each of the layers. If no net energy gain or loss occurs at any of the levels, then the temperatures obtained are the radiative equilibrium values. At the top of the atmosphere we must have energy balance, so that

$$\frac{S_0}{4}\left(1-\alpha_p\right)=\sigma T_e^4=\sigma T_1^4 \qquad (3.47)$$

We thus know immediately that the top layer temperature must equal the emission temperature of the planet, since, in this approximation, the only longwave emission that escapes to space comes from the upper layer. The energy balance at layer 1 is

$$\sigma T_2^4=2\sigma T_1^4 \qquad (3.48)$$

The balance at layer 2 yields

$$\sigma T_1^4+\sigma T_s^4=2\sigma T_2^4 \qquad (3.49)$$

[1]Goody and Walker (1972).

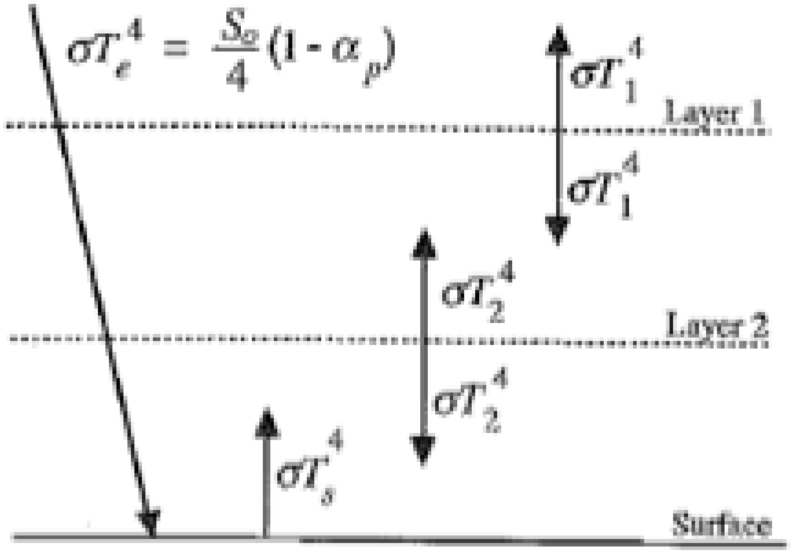


Fig. 3.10 Diagram of simple two-layer radiative equilibrium model for the atmosphere-tem, showing the fluxes of radiant energy.

Hartmann book

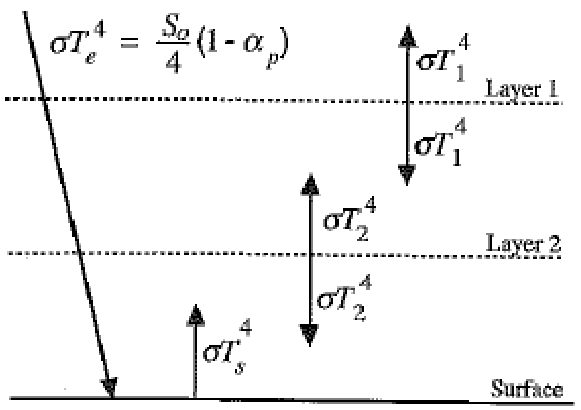


Fig. 3.10 Diagram of simple two-layer radiative equilibrium model for the atmosphere–Earth system, showing the fluxes of radiant energy.

And the balance at the surface is

$$\frac{S_0}{4}\left(1-\alpha_p\right)+\sigma T_2^4=\sigma T_s^4 \tag{3.50}$$

The critical effect of an atmosphere that absorbs and emits longwave radiation appears in (3.50). The energy supplied to the surface by the sun is augmented by a downward flux of longwave radiation from the atmosphere. This allows the surface temperature to rise significantly above the value it would have in the absence of an atmosphere.

We can use (3.47) through (3.50) to solve for the surface temperature.

$$T_s^4=3\frac{(S_0/4)\left(1-\alpha_p\right)}{\sigma}=3T_e^4 \tag{3.51}$$

By extension, if such a model atmosphere has an arbitrary number of layers, n, the surface temperature in equilibrium will be

$$T_s=\sqrt[4]{n+1}\;T_e \tag{3.52}$$

The radiative equilibrium surface temperature for a two-layer atmosphere is 335 K, which is much hotter than Earth's surface temperature. Radiative equilibrium is not a good approximation for the surface temperature, since we know that latent and sensible heat fluxes remove substantial amounts of energy from the surface.

If the atmosphere absorbs no solar radiation, the energy balance for a thin layer of emissivity ε at the top of the atmosphere is between absorption of the flux of terrestrial radiation from below and the emission from the layer itself.

$$\varepsilon\sigma T_e^4=2\varepsilon\sigma T_{\text{strat}}^4 \tag{3.53}$$

where T_{strat} is the temperature at the outer edge of the atmosphere, which we may take to be the stratosphere.

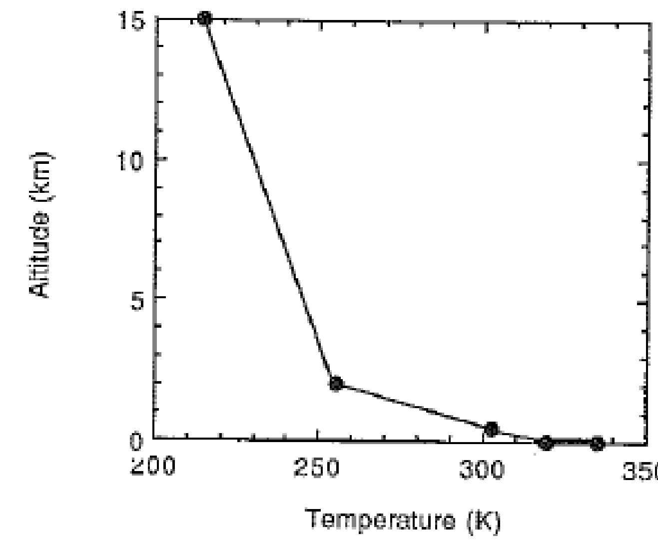


Fig. 3.11 Plot of temperature profile obtained from the simple two-level atmosphere radiative equilibrium model.

A thin layer of atmosphere near the surface absorbs a fraction ε of the emission from above and below and emits in both directions. The temperature of the air adjacent to the surface, T_{SA}, may be derived from the energy balance there.

$$\varepsilon\sigma T_s^4+\varepsilon\sigma T_2^4=2\varepsilon\sigma T_{SA}^4 \tag{3.54}$$

We can solve for all of the temperatures and obtain the following values.

$T_1 = 255$ K	$T_2 = 303$ K	$T_s = 335$ K
$T_{\text{strat}} = 214$ K	$T_{SA} = 320$ K	

These temperatures are plotted in Fig. 3.11. In pure radiative equilibrium the temperatures of the surface and the air in contact with the surface are different. This discontinuity is caused by the absorption of solar radiation at the surface. Such discontinuities are usually greatly suppressed in reality because of efficient heat transport by conduction and convection.

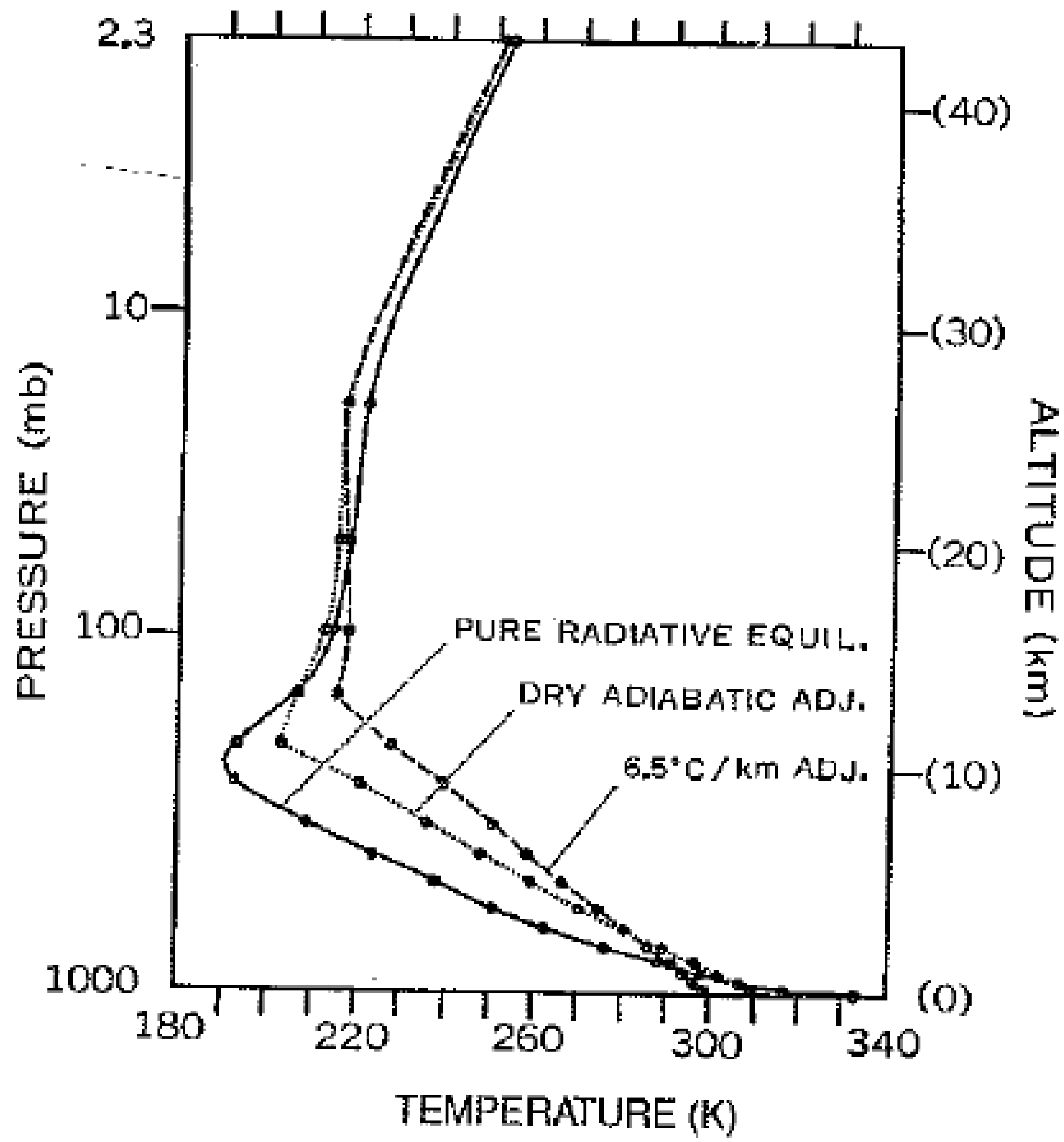
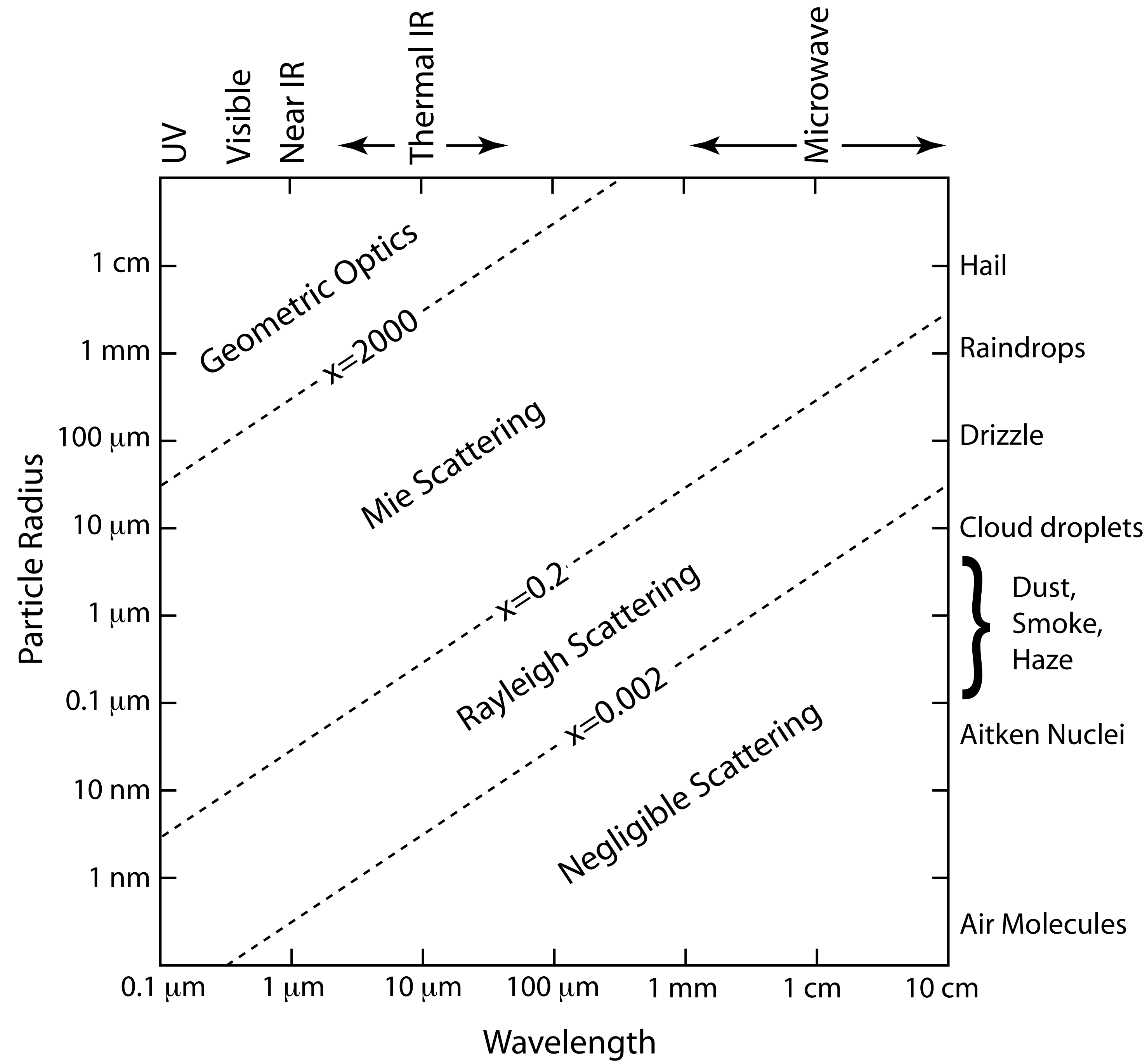


Fig. 3.16 Calculated temperature profiles for radiative equilibrium, and thermal equilibrium with lapse rates of 9.8°C km^{-1} and 6.5°C km^{-1}. [From Manabe and Strickler (1964). Reprinted with permission from the American Meteorological Society.]

table 3.1 Petty

Region	Spectral range	Fraction of solar output	Remarks
X rays	$\lambda < 0.01\ \mu m$		Photoionizes all species; absorbed in upper atmosphere
Extreme UV	$0.01 < \lambda < 0.1\ \mu m$	3×10^{-6}	Photoionizes O_2 and N_2; absorbed above 90 km
Far UV	$0.1 < \lambda < 0.2\ \mu m$	0.01%	Photodissociates O_2; absorbed above 50 km
UV-C	$0.2 < \lambda < 0.28\ \mu m$	0.5%	Photodissociates O_2 and O_3; absorbed between 30 and 60 km
UV-B	$0.28 < \lambda < 0.32\ \mu m$	1.3%	Mostly absorbed by O_3 in stratosphere; responsible for sunburn
UV-A	$0.32 < \lambda < 0.4\ \mu m$	6.2%	Reaches surface
Visible	$0.4 < \lambda < 0.7\ \mu m$	39%	Atmosphere mostly transparent
Near IR	$0.7 < \lambda < 4\ \mu m$	52%	Partially absorbed, mainly by water vapor
Thermal IR	$4 < \lambda < 50\ \mu m$	0.9%	Absorbed and emitted by water vapor, carbon dioxide, ozone, and other trace gases
Far IR	$0.05 < \lambda < 1$ mm		Absorbed by water vapor
Microwave	$\lambda > 1$ mm		Clouds semi-transparent


UV
Visible
Near IR
Thermal IR
Microwave
1 cm
1 mm
100 μm
10 μm
1 μm
0.1 μm
10 nm
1 nm
Particle Radius
Hail
Raindrops
Drizzle
Cloud droplets
Dust,
Smoke,
Haze
Aitken Nuclei
Air Molecules
Geometric Optics
x=2000
Mie Scattering
x=0.2
Rayleigh Scattering
x=0.002
Negligible Scattering
0.1 μm
1 μm
10 μm
100 μm
1 mm
1 cm
10 cm
Wavelength

homework (last one :)) due after Thanksgiving

WH 4.41

WH 4.42

Problem 2.14: A typical laser pointer used in lectures puts out 5 mW of power into a nearly parallel beam with a diameter of 5 mm. (a) What is the *flux density* normal to the beam, and how does it compare with the typical clear-sky solar flux (at ground level) on a surface normal to the beam of 1000 W m^{-2}? (b) If beam can be assumed to be confined to a cone of angular diameter 1 milliradian, what is the *intensity* of the beam in watts per steradian, and how does this compare with the intensity of sunlight computed from the above solar flux and an angular diameter of 0.5° for the sun's disk?