Lecture 3: Atmospheric Radiative Transfer and Climate

Chapter 3: Atmospheric Radiative Transfer and Climate

- ❑ Solar and infrared radiation
- ❑ selective absorption and emission
- ❑ Selective absorption and emission
- ❑ Cloud and radiation
- ❑ Radiative-convective equilibrium

Importance of Radiation Transfer

- Virtually all the exchange of energy between the Earth and the rest of the universe takes place by radiation transfer.

- Radiation transfer is also a major way of energy transfer between the atmosphere and the underlying surface and between different layers of the atmosphere.

Radiation Intensity and Wavelength

$$E_v = hv$$

frequency

Planck's constant

- The shorter the wavelength of the radiation, the larger the amount of energy carried by that radiation.

Spectrum of Radiation

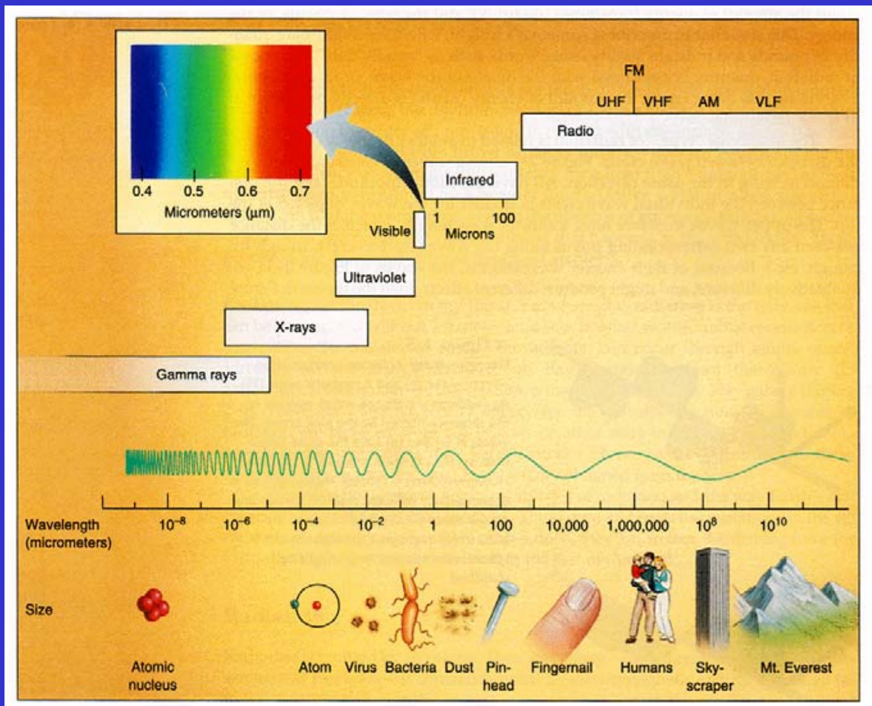


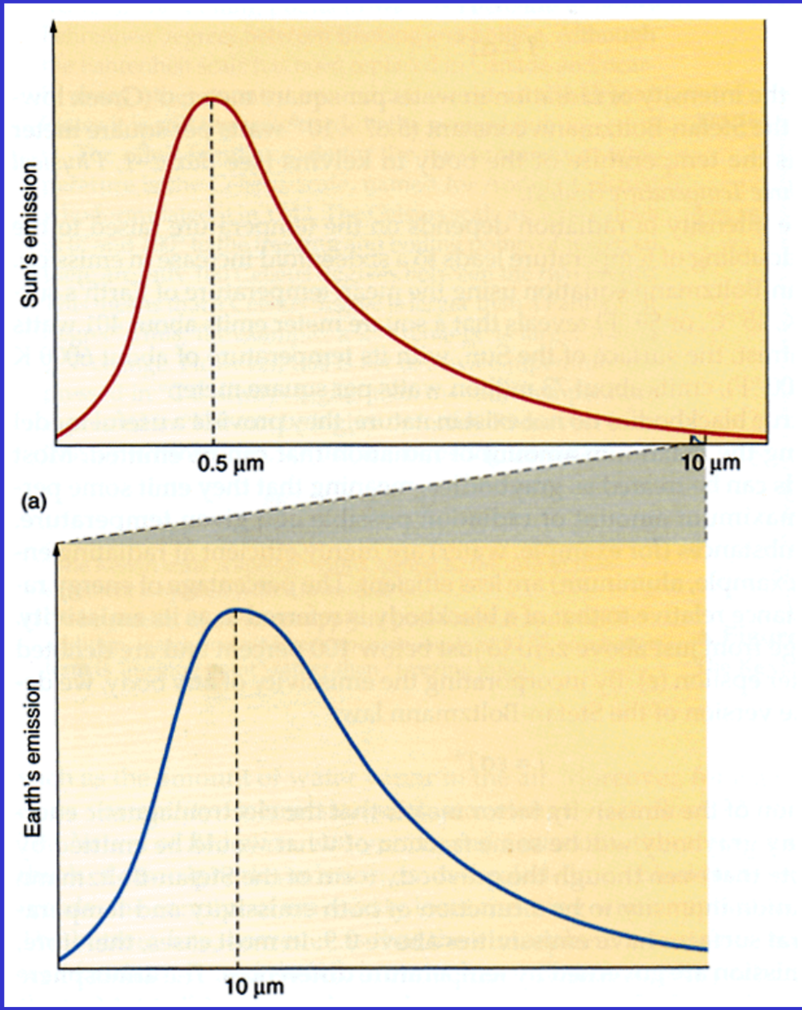
Table 2–1 Wavelength Categorizations

Type of Energy	Wavelength (micrometers)
Gamma	<0.0001
X ray	0.0001 to 0.01
Ultraviolet	0.01 to 0.4
Visible	0.4 to 0.7
Near Infrared (NIR)	0.7 to 4.0
Thermal Infrared	4 to 100
Microwave	100 to 1,000,000 (1 meter)
Radio	>1,000,000 (1 meter)

(from *Understanding Weather & Climate*)

- ❑ Radiation energy comes in an infinite number of wavelengths.
- ❑ We can divide these wavelengths into a few bands.

Solar and Terrestrial Radiation



(from *Understanding Weather & Climate*)

- All objectives radiate energy, not merely at one single wavelength but over a wide range of different wavelengths.
- The sun radiates more energy than the Earth.
- The greatest intensity of solar energy is radiated at a wavelength much shorter than that of the greatest energy emitted by the Earth.

Stefan-Boltzmann Law

$$E = \sigma T^4$$

E = radiation emitted in W/m^2

σ= 5.67 x 10^{-8} W/m^2 * K *sec

T = temperate (K)

- The single factor that determine how much energy is emitted by a blackbody is its temperature.
- The intensity of energy radiated by a blackbody increases according to the fourth power of its absolute temperature.
- This relationship is called the Stefan-Boltzmann Law.

Apply Stefan-Boltzmann Law To Sun and Earth

- **Sun**

  $E_s = (5.67 \times 10^{-8}\ W/m^2\ K^4) * (6000K)^4$

  $= 73{,}483{,}200\ W/m^2$

- **Earth**

  $E_e = (5.67 \times 10^{-8}\ W/m^2\ K^4) * (300K)^4$

  $= 459\ W/m^2$

- Sun emits about 160,000 times more radiation per unit area than the Earth because Sun's temperature is about 20 times higher than Earth's temperature.

  ➔ $20^4 = 160{,}000$

Wien's Law

$$\lambda_{max} = w/T$$

λ_{max} = wavelength (micrometers)

W = 2897 μm K

T = temperate (K)

- Wien's law relates an objective's maximum emitted wavelength of radiation to the objective's temperature.
- It states that the wavelength of the maximum emitted radiation by an object is inversely proportional to the objective's absolute temperature.

Apply Wien's Law To Sun and Earth

- **Sun**

  λ_{max} = 2898 μm K / 6000K

  = 0.483 μm

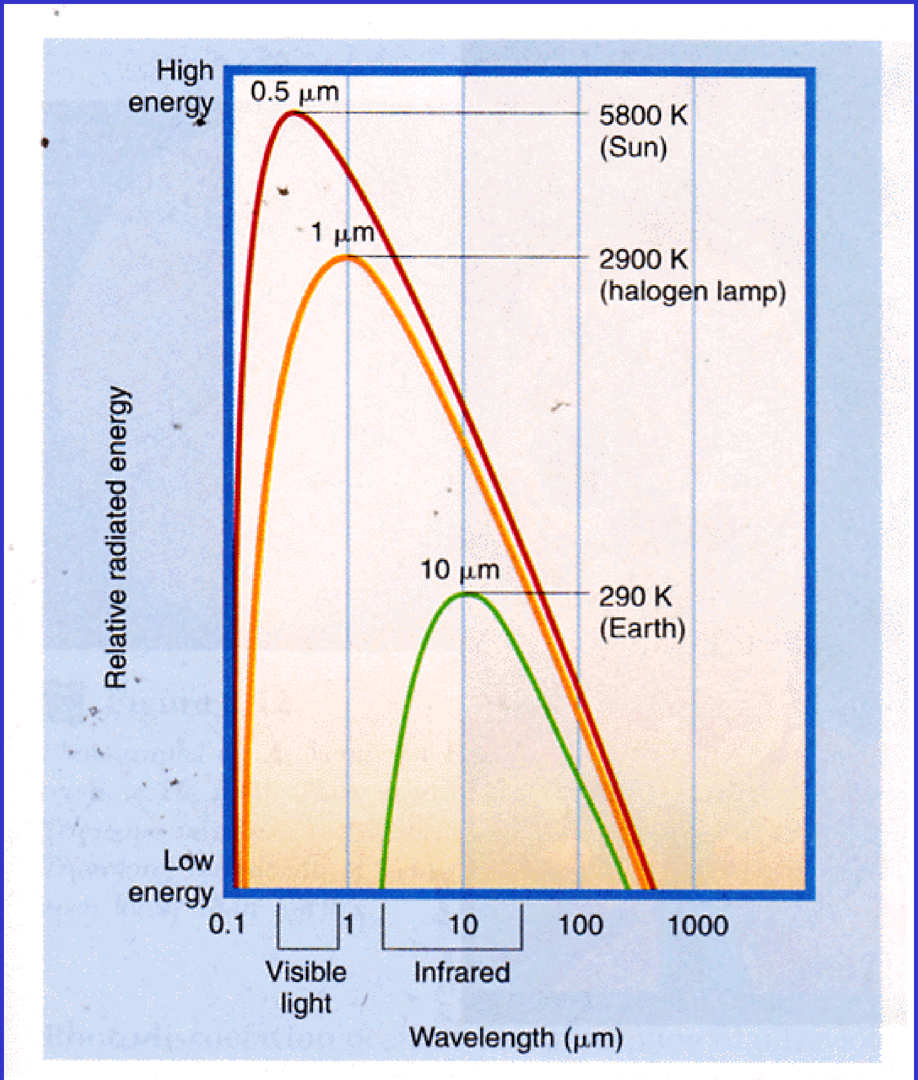
- **Earth**

  λ_{max} = 2898 μm K / 300K

  = 9.66 μm

- Sun radiates its maximum energy within the visible portion of the radiation spectrum, while Earth radiates its maximum energy in the infrared portion of the spectrum.

Wavelength and Temperature

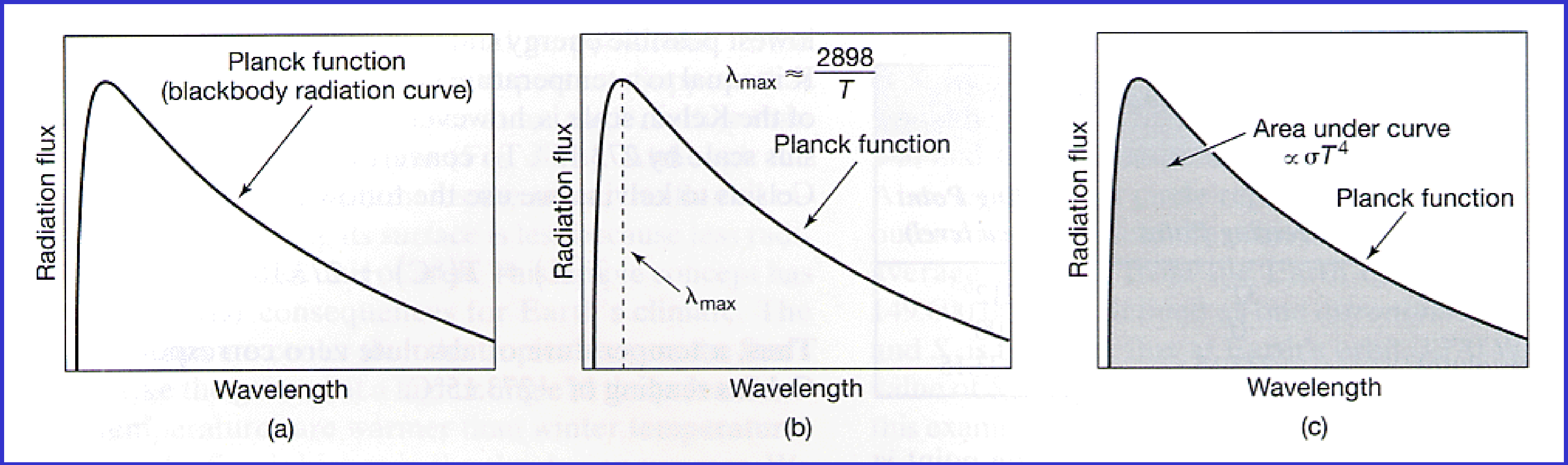


(from *Meteorology: Understanding the Atmosphere*)

- The hotter the objective, the shorter the wavelength of the peak radiation.

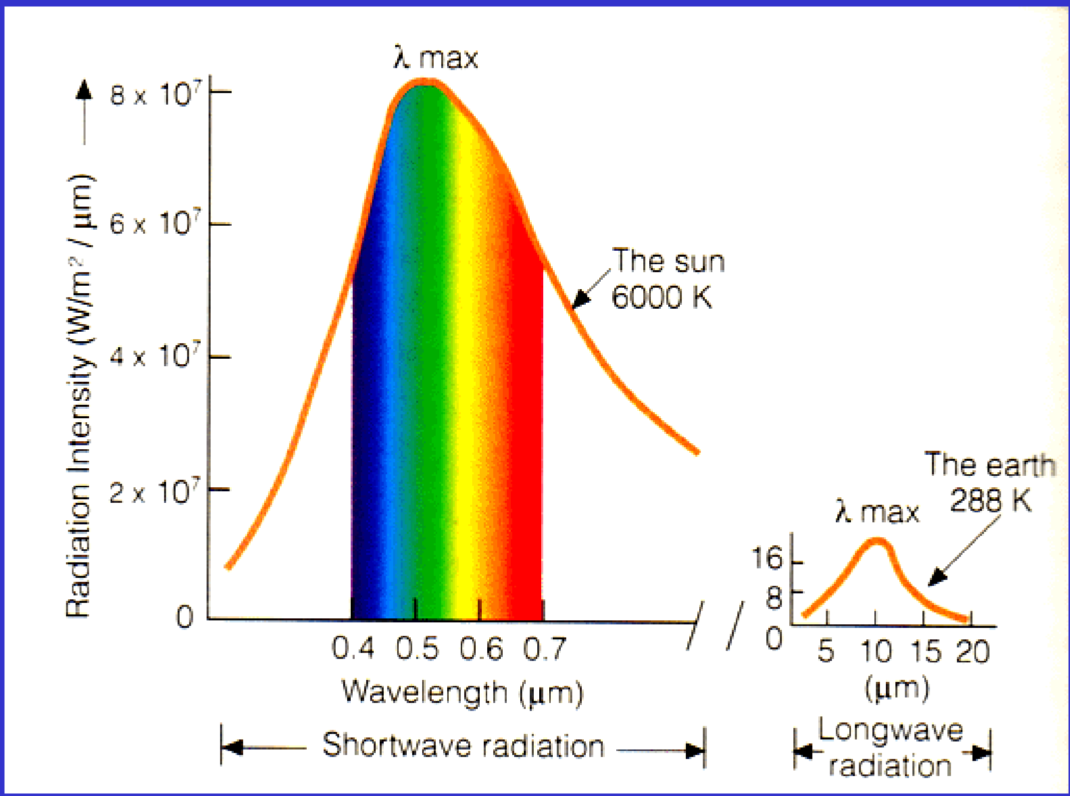
Planck Function

- The Planck function relates the intensity of radiation from a blackbody to its wavelength.



(from *The Earth System*)

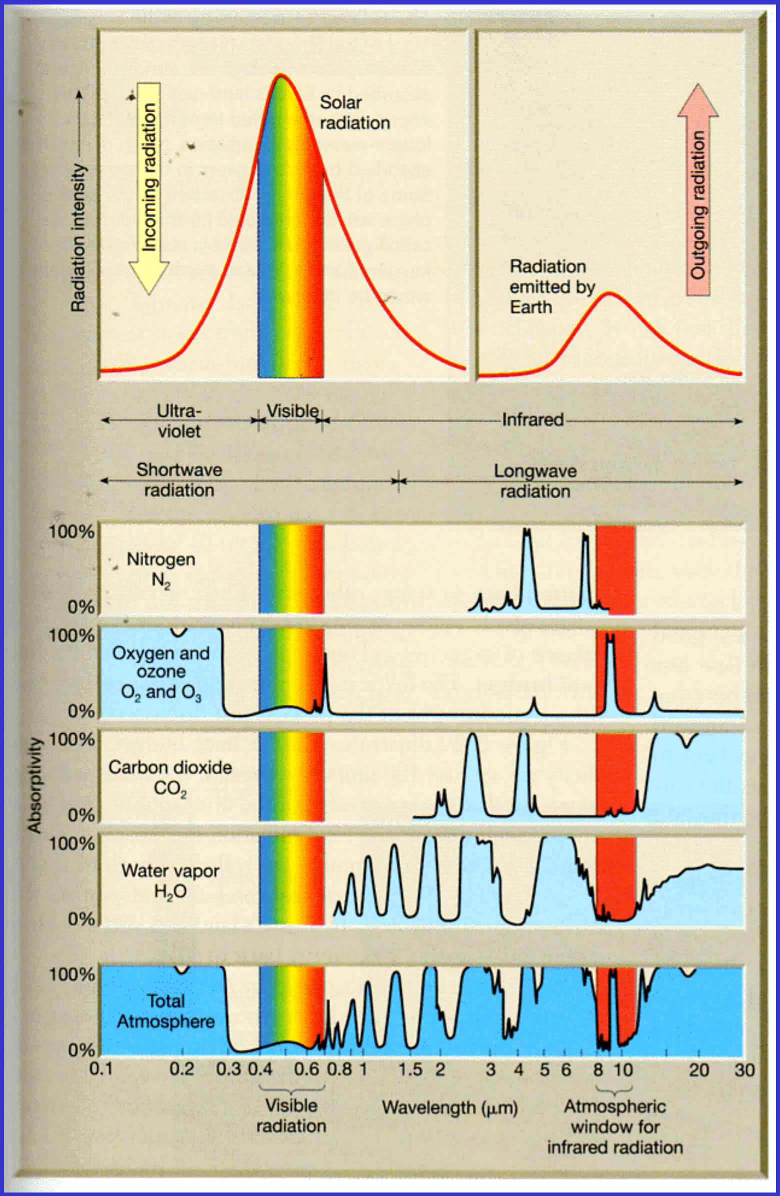
Shortwave and Longwave Radiations



(from *Meteorology: Understanding the Atmosphere*)

- ❑ Solar radiation is often referred to as "shortwave radiation".
- ❑ Terrestrial radiation is referred to as "longwave radiation".

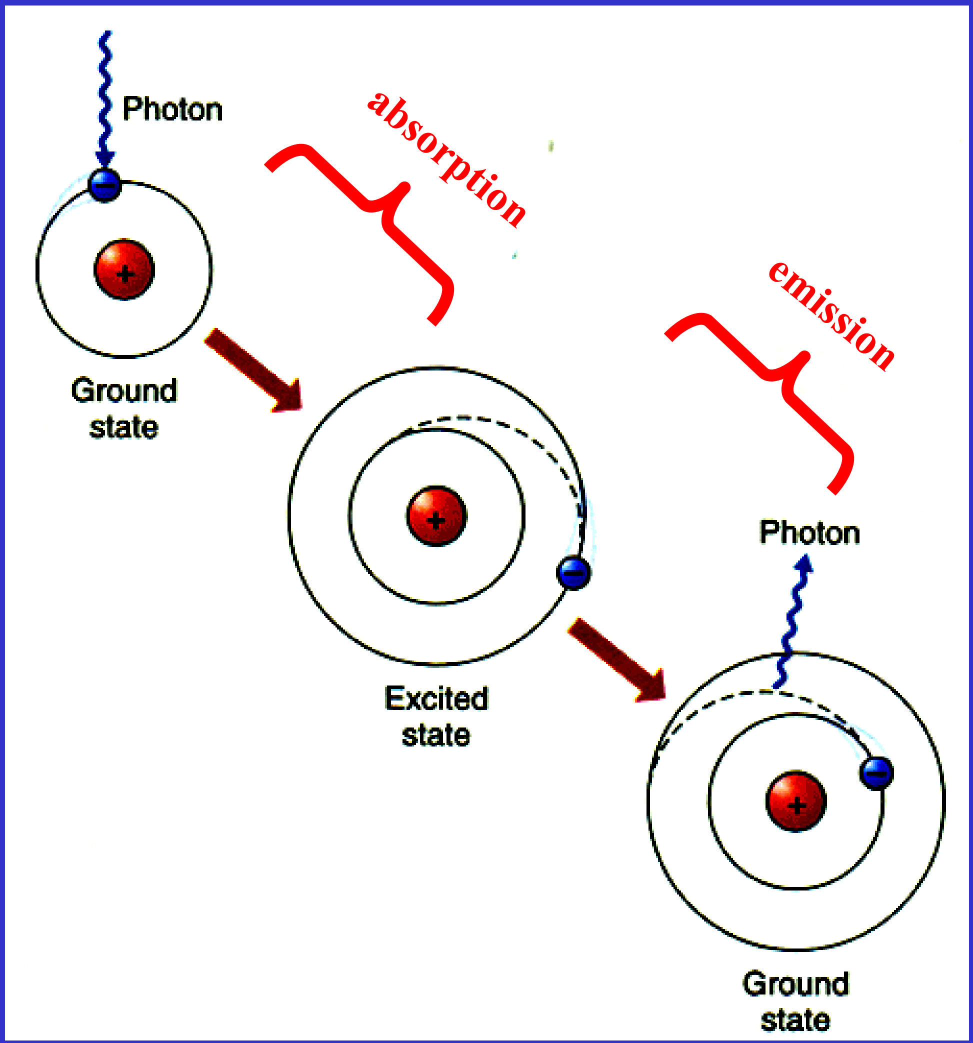
Selective Absorption and Emission



- The atmosphere is not a perfect blackbody, it absorbs some wavelength of radiation and is transparent to others (such as solar radiation). ➔ Greenhouse effect.
- Objective that selectively absorbs radiation usually selectively emit radiation at the same wavelength.
- For example, water vapor and CO2 are strong absorbers of infrared radiation and poor absorbers of visible solar radiation.

(from *The Atmosphere*)

Why Selective Absorption/Emission?

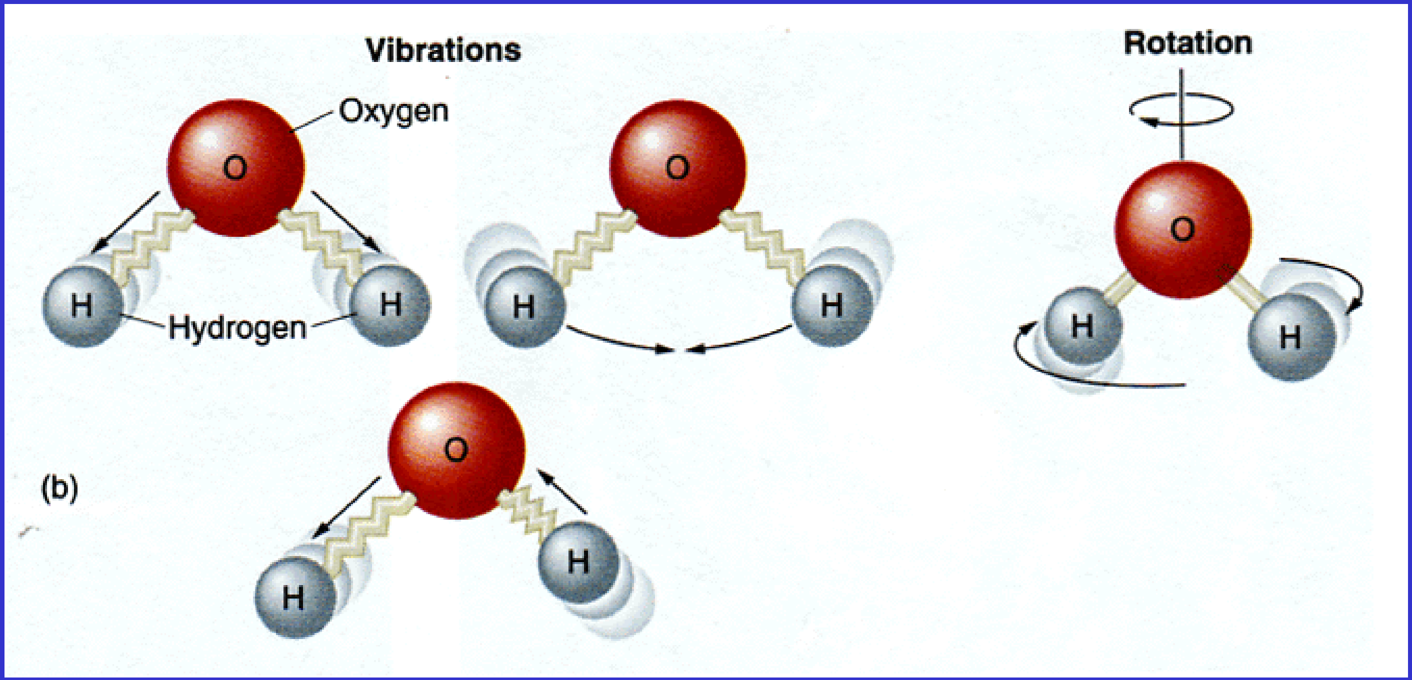


(from *Understanding Weather & Climate*)

- Radiation energy is absorbed or emitted to change the energy levels of atoms or molecular.
- The energy levels of atoms and molecular are discrete but not continuous.
- Therefore, atoms and molecular can absorb or emit certain amounts of energy that correspond to the differences between the differences of their energy levels.

➔ Absorb or emit at selective frequencies.

Different Forms of Energy Levels

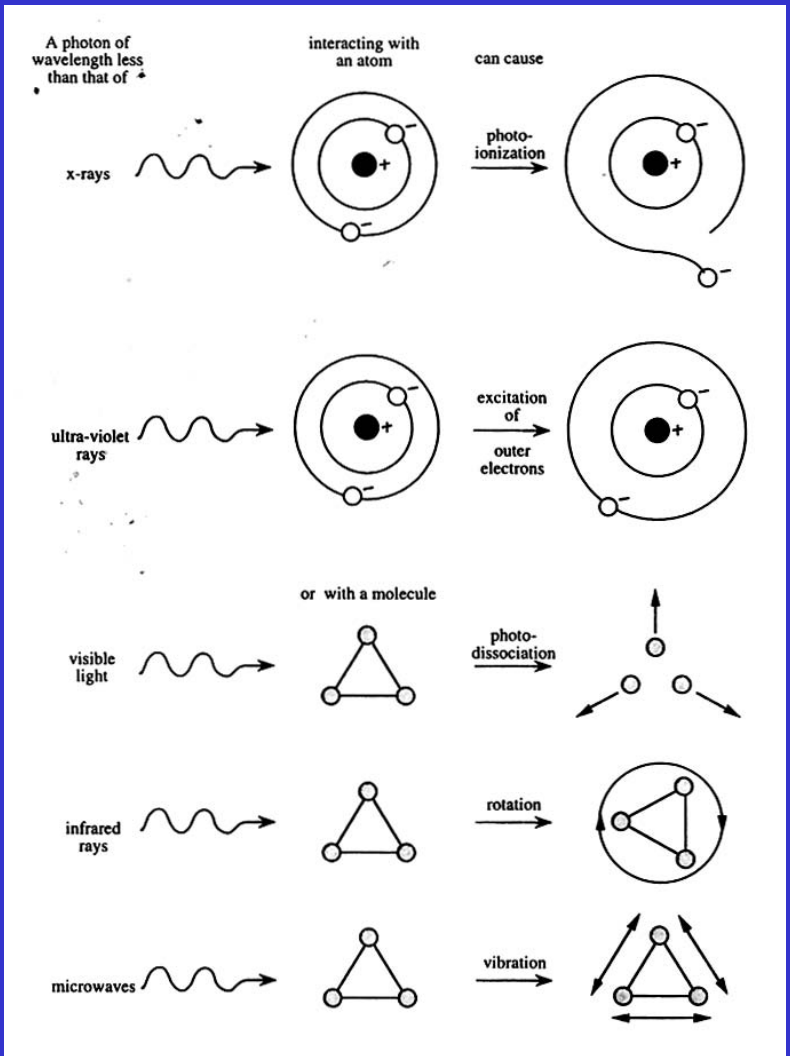


(from *Understanding Weather & Climate*)

- The energy of a molecule can be stored in (1) translational (the gross movement of molecules or atoms through space), (2) vibrational, (3) rotational, and (4) electronic (energy related to the orbit) forms.

Energy Required to Change the Levels

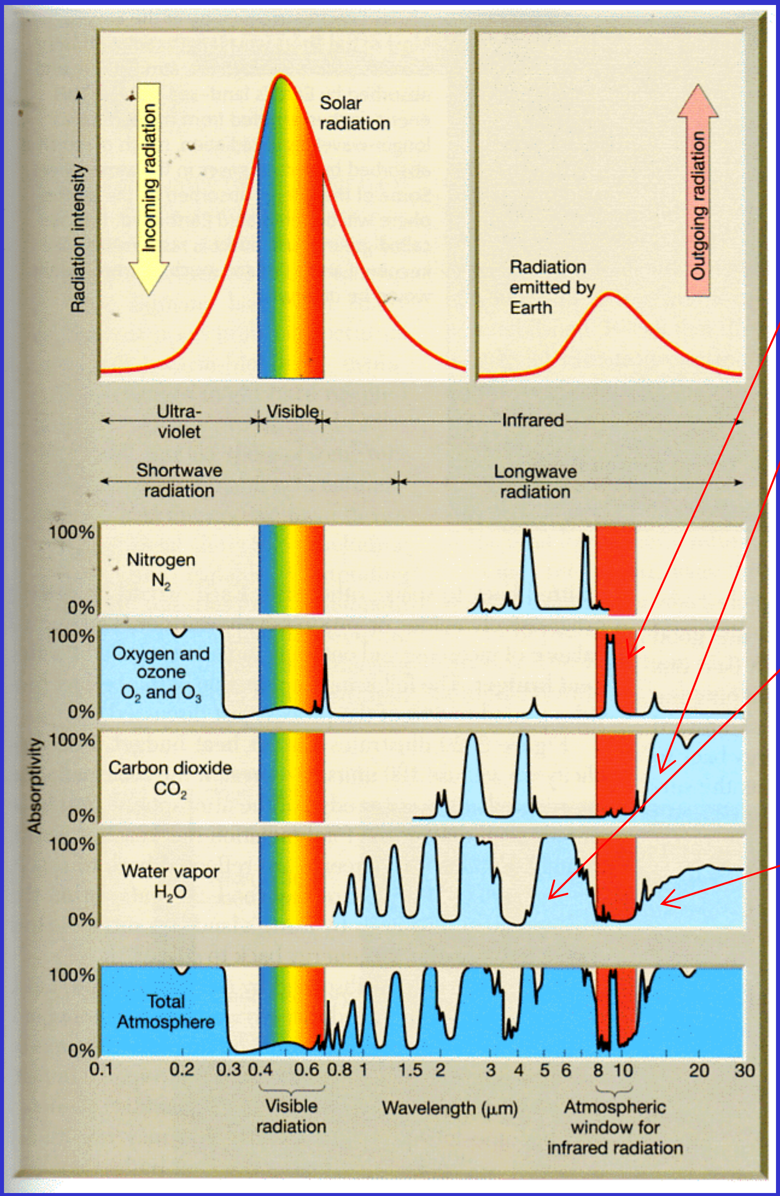
$$E_{\text{Total}} = E_{\text{Translational}} + E_{\text{Rotational}} + E_{\text{Vibrational}} + E_{\text{Electronic}}$$



(from *Is The Temperature Rising?*)

- Translational energy corresponds to the gross movement of molecules or atoms through space and is not quantized.
- For tiny objects such as molecules in the atmosphere, the energy of rotation is quantized and can take on only discrete values.
- Molecular energy can be stored in the vibrations about the stable bonding of atmos.
- The most energetic photons (with shortest wavelength) are at the top of the figure, toward the bottom, energy level decreases, and wavelengths increase.

Absorption of Terrestrial Radiation



(from *The Atmosphere*)

In the middle of the atmospheric window sits the 9.6- µm band of ozone absorption.

The bending mode of CO2 produces a very strong vibration–rotation absorption band near 15 µm.

Water vapor has an important vibration–rotation band near 6.3 µm.

Water vapor has pure rotational lines strongly absorbs terrestrial emission at wavelengths greater than about 12 µm.

The key absorption features for terrestrial radiation are (1) a water-vapor vibration–rotation band near 6.3 mm, (2) the 9.6-mm band of ozone, (3) the 15-mm band of carbon dioxide, and (4) the dense rotational bands of water vapor that become increasingly important at wavelengths longer than 12 mm.

Infrared absorption spectra for various atmospheric gases

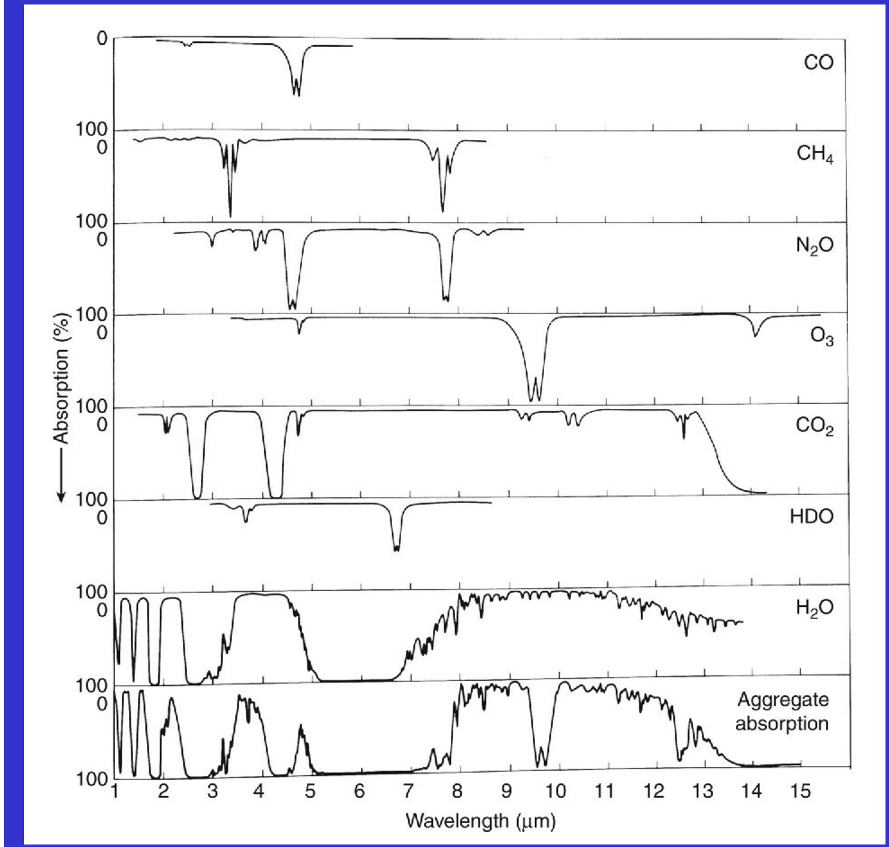


TABLE 3.1 Wavelengths of Vibrational Modes of Some Important Atmospheric Molecules

Species	Vibrational modes		
	ν1	ν2	ν3
CO	4.67		
CO_2		15.0	4.26
N_2O	7.78	17.0	4.49
H_2O	2.73	6.27	2.65
O_3	9.01	14.2	9.59
NO	5.25		
NO_2	7.66	13.25	6.17
CH_4	3.43	6.52	3.31
CH_4	5.25		

Units are in microns (μm).
From Herzberg and Herzberg © 1957 from McGraw Hill, Inc. and Shimanouchi, 1967a,b, 1968.

(from Global Physical Climatology)

- ❑ It is the minor trace concentrations of polyatomic molecules that determine the infrared transmissivity of the atmosphere.
- ❑ The most important gases are water vapor, carbon dioxide, and ozone (in that order), but many other gases contribute significantly

Absorption of Solar Radiation in the Atmosphere

- Visible radiation is too energetic to be absorbed by most of the gases in the atmosphere and not energetic enough to photodissociate them, so that the atmosphere is almost transparent to it.
- Solar radiation with wavelengths between about 0.75 mm and 5 mm, which we will call near-infrared radiation, is absorbed weakly by water, carbon dioxide, ozone, and oxygen.
- Most of the ultraviolet radiation from the Sun with wavelengths shorter than 0.2 mm is absorbed in the upper atmosphere through the photodissociation and ionization of nitrogen and oxygen.
- Radiation at frequencies between 0.2 mm and 0.3 mm is absorbed by ozone in the stratosphere.

Units of Wavelength

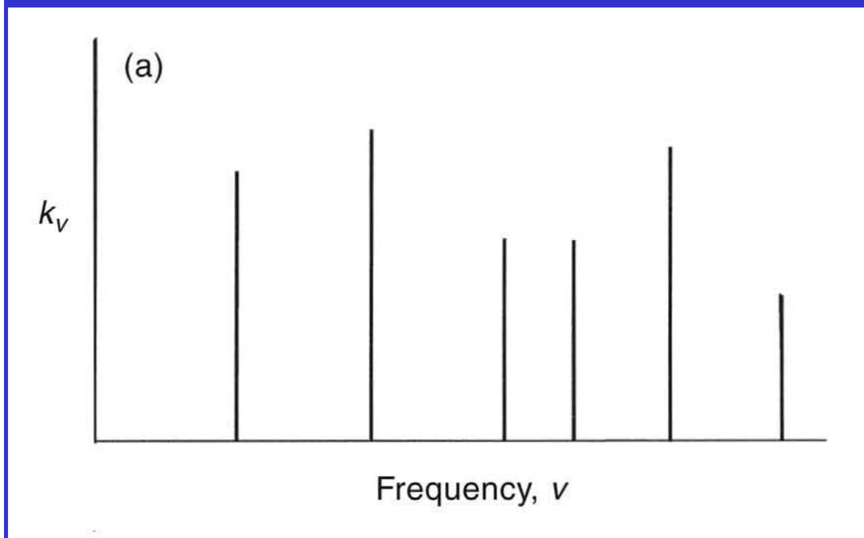
$$\boxed{mm = 10^{3}\ \mu m = 10^{6}\ nm = 10^{-3}\ m}$$

mm: millimeter

µm: micrometer

nm: nanometer

Absorption Lines and Bands

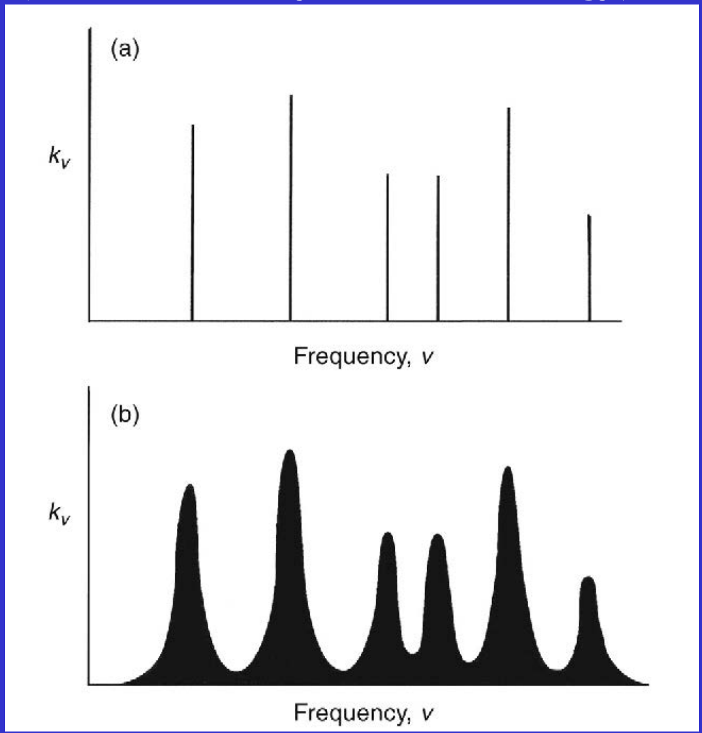


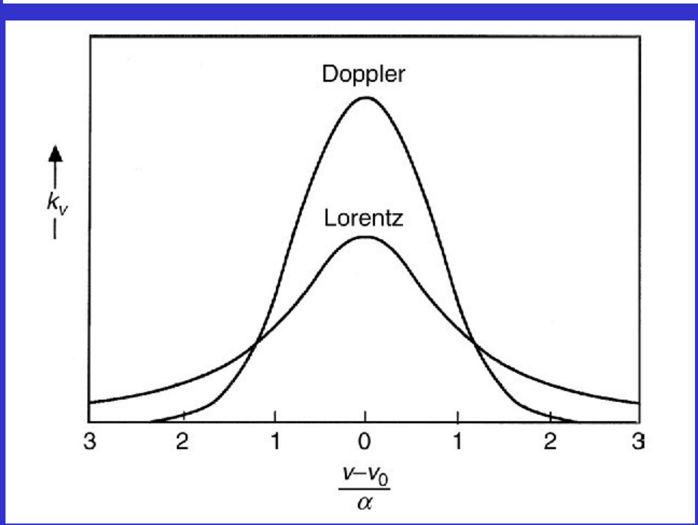
(from *Global Physical Climatology*)

- Air molecular absorption take place at the discrete frequencies corresponding to an energy transition of an atmospheric gas.
- **Absorption line**: Each of these discrete absorption features an absorption line.
- **Absorption band:** The collection of such absorption lines in a particular frequency interval can be called an absorption band.
- water vapor has many rotational absorption lines at closely spaced frequencies, which form a rotation band that absorbs much of Earth's emission at wavelengths between 12 mm and 200 mm.

Lines Broadening

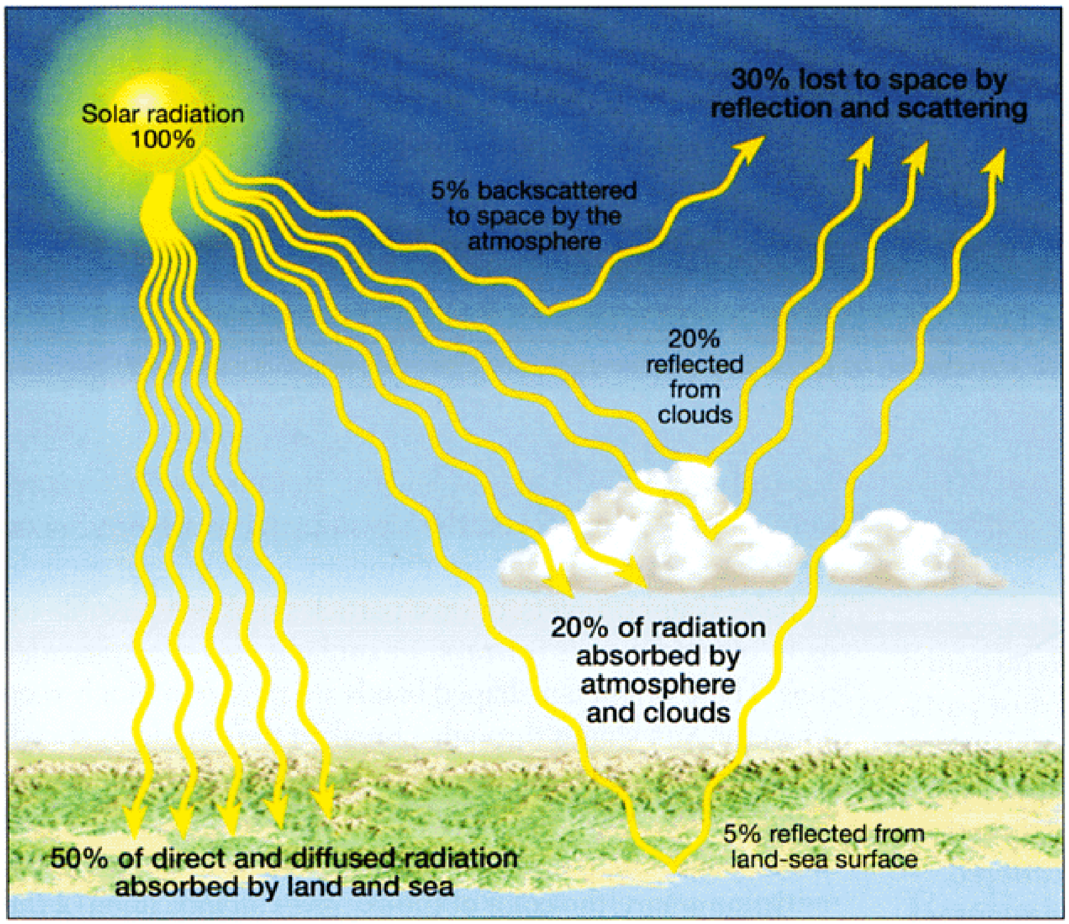
(from *Global Physical Climatology*)





- Air molecular absorption takes place at the discrete frequencies, but the actual spectra are characterized by absorption "bands".
- **Natural Broadening**: associated with the finite time of photon emission or absorption and with the uncertainty principle. This mechanism is usually less important than pressure or Doppler broadening.
- **Doppler Broadening**: The atoms in a gas which are emitting radiation will have a distribution of velocities. Each photon emitted will be "red"- or "blue"-shifted by the Doppler effect depending on the velocity of the atom relative to the observer. (Doppler profile; important at high altitudes).
- **Pressure/collision Broadening**: The collision of other particles with the emitting particle interrupts the emission process, and by shortening the characteristic time for the process, increases the uncertainty in the energy emitted. (Lorentz profile; important in the troposphere)

Absorption, Reflection, Scattering



(from *The Atmosphere*)

- What happens to incoming solar radiations?

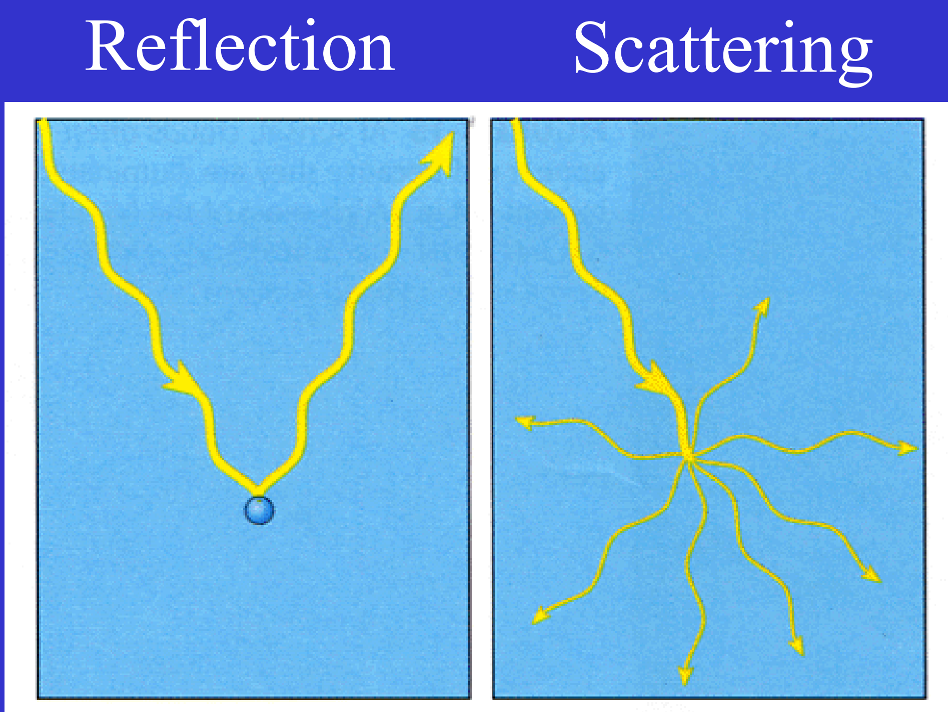
(1) Absorption

(2) Reflection

(3) Scattering

(4) Transmission (through the atmosphere)

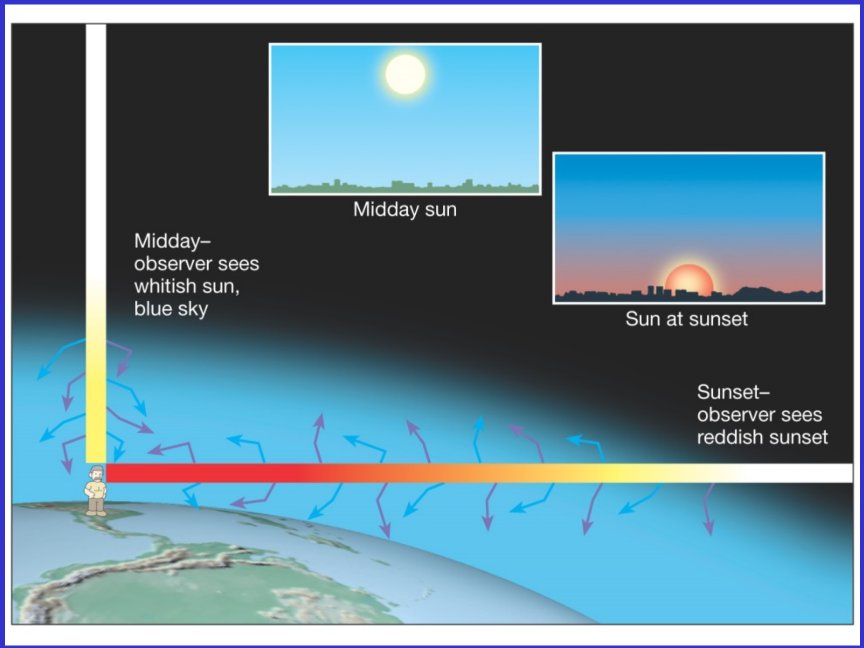
Reflection and Scattering



(from *The Atmosphere*)

- Reflection: light bounces back from an objective at the same angle at which it encounters a surface and with the same intensity.
- Scattering: light is split into a larger number of rays, traveling in different directions.
- Although scattering disperses light both forward and backward (backscattering), more energy is dispersed in the forward direction.

Scattering and Colors



- Short wavelengths (blue and violet) of visible light are scattered more effectively than are longer wavelengths (red, orange). Therefore, when the Sun is overhead, an observed can look in any direction and see predominantly blue light that was selectively scattered by the gases in the atmosphere.
- At sunset, the path of light must take through the atmosphere is much longer. Most of the blue light is scattered before it reaches an observer. Thus the Sun appears reddish in color.

Radiant Intensity (Irradiance)

The magnitude of the ***radiant intensity*** (Iν) is given in energy per unit time, per unit area, per unit of frequency interval, per unit of solid angle.

$$dF_v = I_v \cos\theta \, d\omega \, dA \, dv \, dt$$

We concerned mostly with the total energy per unit frequency passing across a unit area of a plane surface from one side to the other.

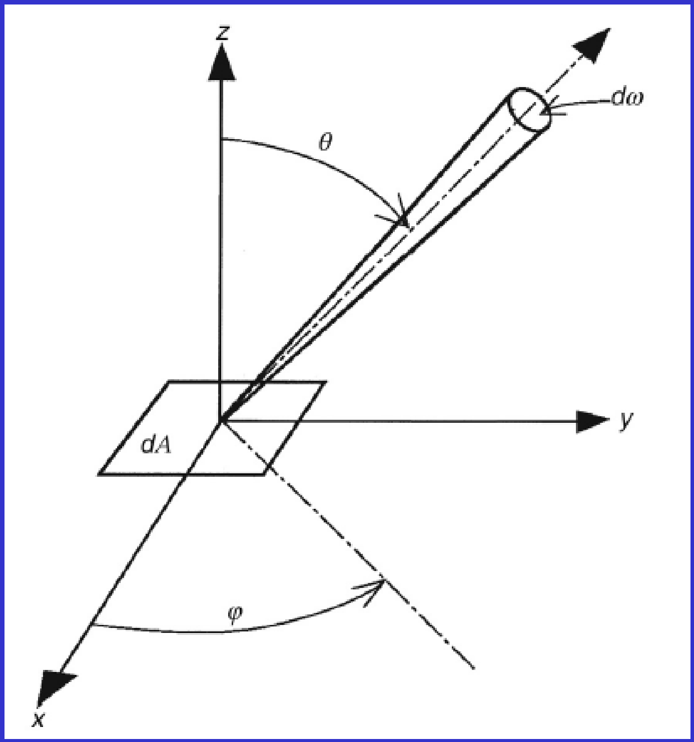
Spectral irradiance

$$F_v = \int_0^{2\pi} \int_0^{\pi/2} I_v(\theta,\psi) \cos\theta \sin\theta \, d\theta \, d\varphi$$

If Fν is integrated over all frequencies, we obtain the irradiance, which has units of watts per meter squared.

$$F = \int_0^{\infty} F_v \, dv$$

Irradiance



$$d\omega = \sin\theta \, d\theta \, d\varphi$$

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Lambert–Bouguer–Beer Law of Extinction

The change in the irradiance (dF) along a path of length ds, where the density of the absorber is ρ_a and the absorption coefficient is k_{abs}, may be written

$$dF = -k_{abs}\rho_a F ds \tag{3.12}$$

In (3.12), F and ds are both measured positive downward. F and k_{abs} depend on frequency, but we have dropped the frequency subscript for economy. The units of k_{abs} in (3.12) must be $m^2\ kg^{-1}$. Because its units are area per unit mass, k_{abs} is sometimes also called the absorption cross-section of the gas in question. From Fig. 3.7, the path length is related to altitude according to

$$dz = -\cos\theta\, ds \tag{3.13}$$

Therefore, (3.12) becomes

absorption constant

$$\cos\theta \frac{dF}{dz} = k_{abs}\rho_a F \tag{3.14}$$

We can define the optical depth (τ) along a vertical path.

$$\boxed{\tau = \int_z^\infty k_{abs}\rho_a\, dz} \tag{3.15}$$

optical depth

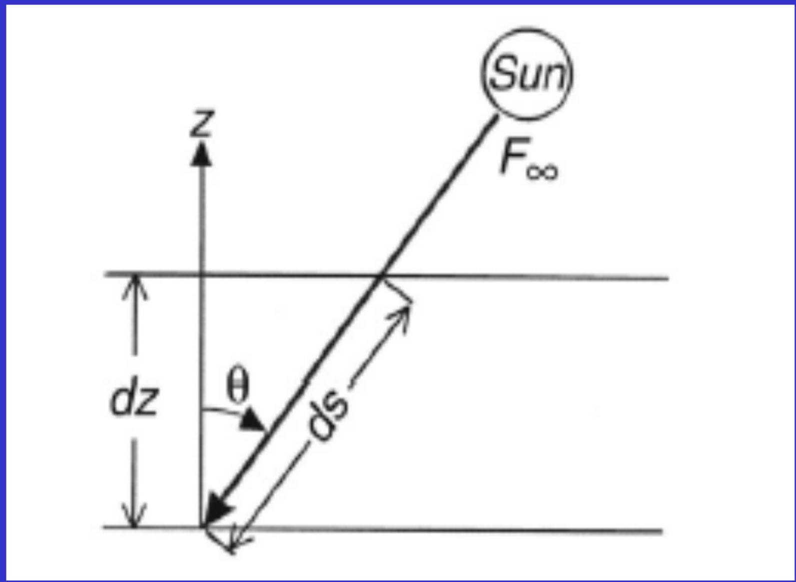
Note that (3.15) implies that $d\tau = -k_{abs}\rho_a dz$, so that we can write (3.14) as

$$\cos\theta \frac{dF}{d\tau} = -F \tag{3.16}$$

This equation has a very simple solution,

$$\boxed{F = F_\infty e^{-\tau/\cos\theta}} \tag{3.17}$$

where F_∞ is, in this case, the downward irradiance at the top of the atmosphere. Thus, the incident radiation decays exponentially along the slant path ds where the optical depth is given by $\tau/\cos\theta$.



- This law states that absorption is linear in the intensity of radiation and the absorber amount.
- The absorption by a layer of depth dz is proportional to the irradiance (F) times the mass of absorber along the path the radiation follows.

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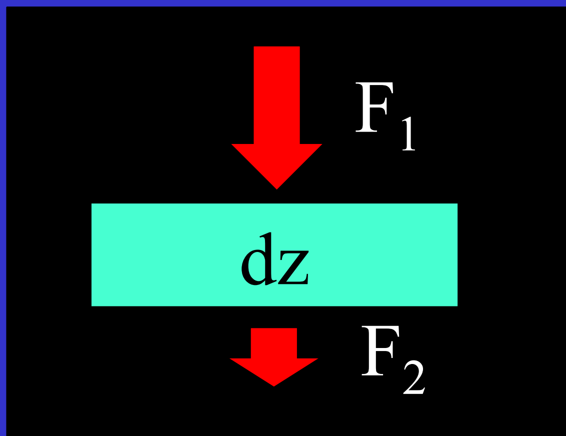
Absorption Rate and Heating Rate



downward irradiance at the TOA

optical depth

$$\text{Absorption rate} = \frac{dF}{dz} = \frac{k_{abs}\rho_a F}{\mu} = -\frac{d\tau}{dz}\frac{F}{\mu} = \frac{F_\infty}{\mu}e^{-\tau/\mu}\frac{\tau}{H}$$

cosθ

scale height

$$\text{Heating rate} = \left.\frac{\partial T}{\partial t}\right|_{rad} = \frac{1}{c_p\rho}\frac{\partial F}{\partial z} = \frac{k_{abs}M_a}{c_p\mu}F$$

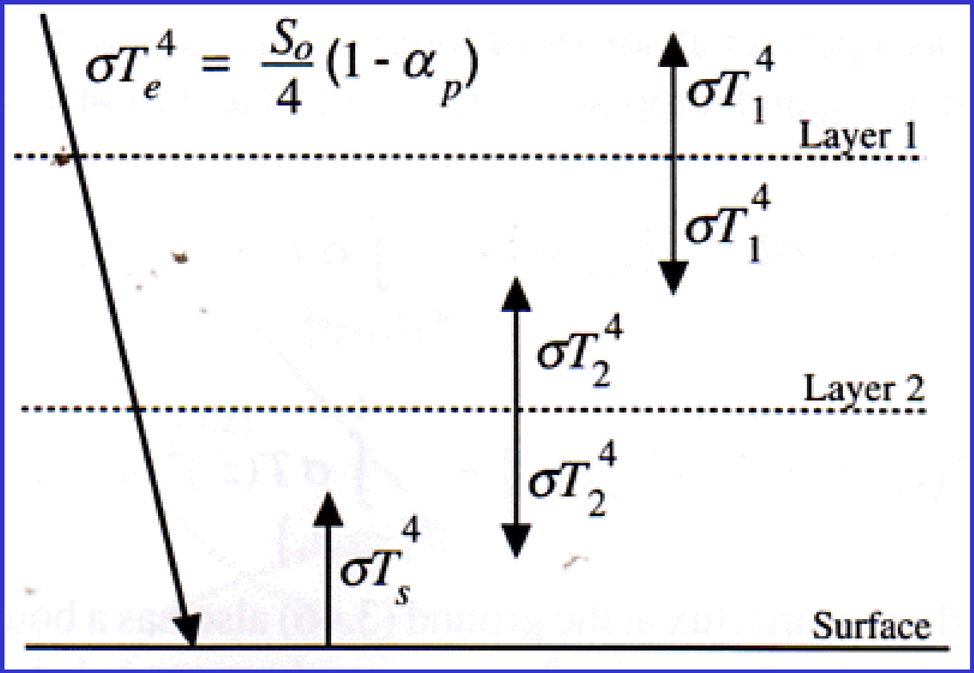
mixing ratio of the absorber

- The vertical distribution of Ma determines where the maximum heating rate should be produced: (1) at the outer extremity of the atmosphere for O2 (constant Ma), (2) in the stratosphere for O3.

3.7 INFRARED RADIATIVE TRANSFER EQUATION: ABSORPTION AND EMISSION

(read this Section 3.7 if you are interested in the details of infrared radiative transfer in the atmosphere)

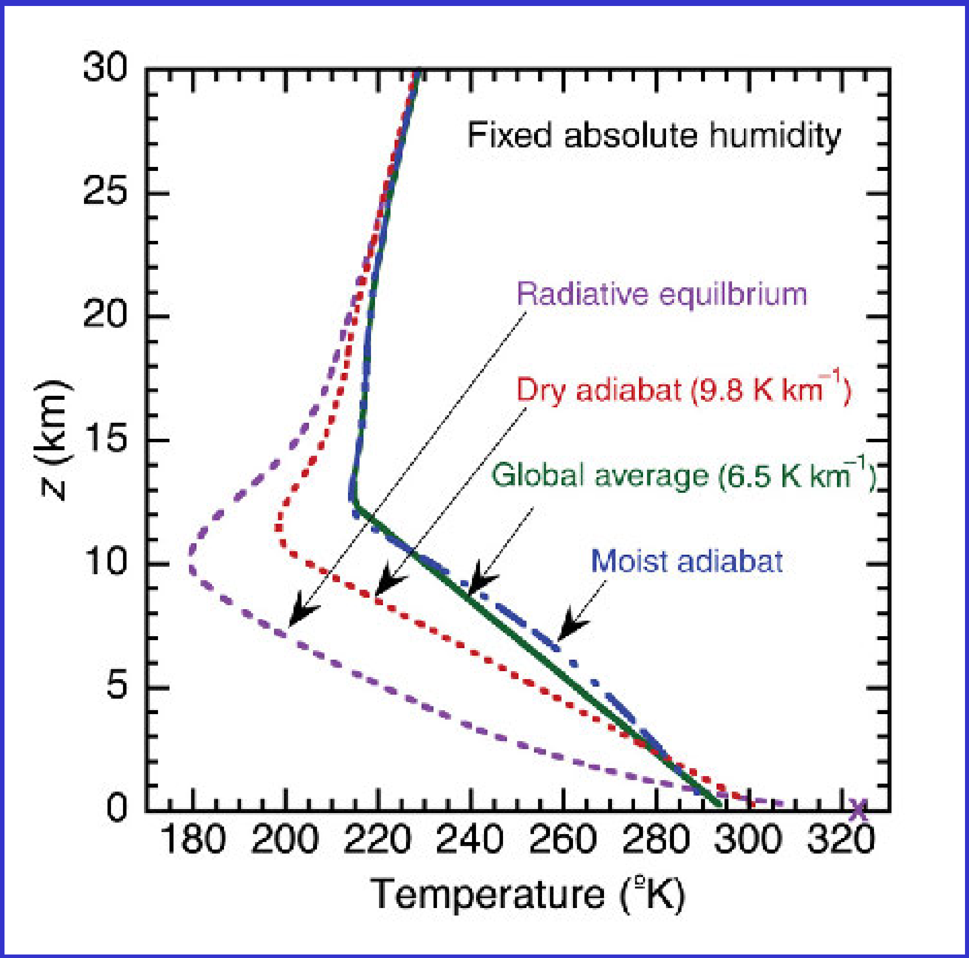
Model of Radiative Equilibrium



(from *Global Physical Climatology*)

- We assume the atmosphere is opaque for longwave radiation and transparent to shortwave radiation.
- We divide the atmosphere into many layers.
- We assume energy is balance at each atmospheric layers.
- We can determine the temperature of each atmospheric layers.

Radiative Equilibrium Temperature



(from *Global Physical Climatology*)

- ❑ The radiative equilibrium temperature calculated from the energy balance model is hydrostatically unstable. (meaning the lapse rate is larger than the dry adiabatic lapse rate).
- ❑ As a result, convections occur.
- ➔ The atmosphere becomes stable with a radiative-convective equilibrium temperature.
- ❑ The global mean temperature profile of Earth's atmosphere is not in radiative equilibrium, but rather in radiative–convective equilibrium.

Clouds and Radiation

- Water droplets and ice particles in clouds have substantial interactions with both solar and terrestrial radiation.
- The nature of these interactions depends on (1) the total mass of water, (2) the size and shape of the droplets or particles, and (3) their distribution in space.
- The problem is often simplified by assuming that clouds are uniform and infinite in the horizontal, which is called **the plane-parallel cloud assumption**.
- If the droplet size distribution and the vertical distribution of humidity are assumed, then the cloud albedo and absorption depend on the total liquid water content of the cloud and the solar zenith angle.
- Cloud liquid water content is defined as the total mass of cloud water in a vertical column of atmosphere per unit of surface area.

Important Roles of Clouds In Global Climate

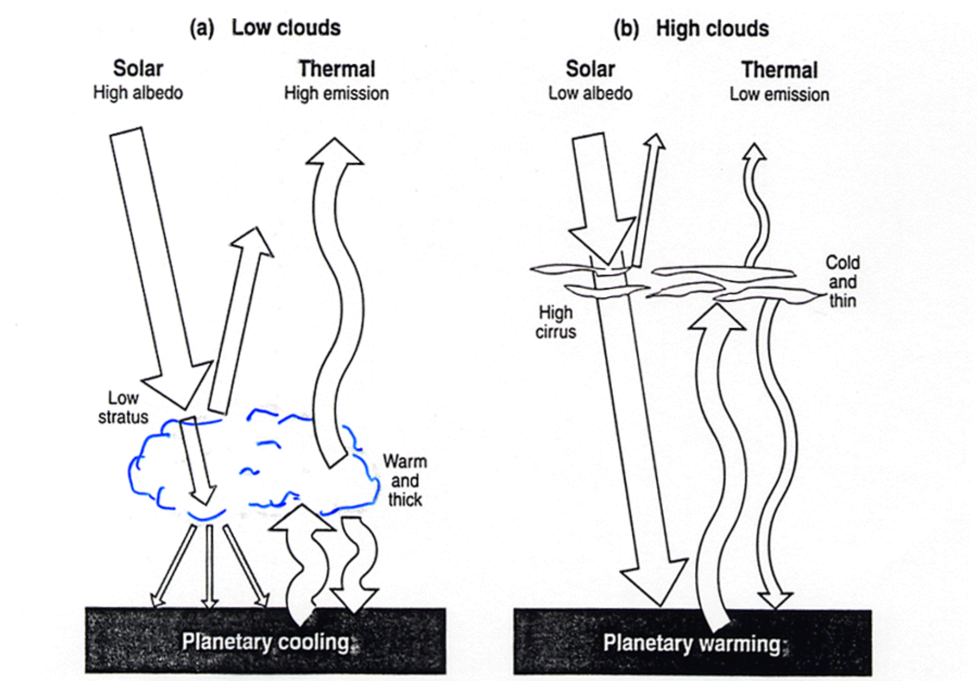


Figure 11.13 The effects of clouds on the flow of radiation and energy in the lower atmosphere and at the surface. Two cases are shown: (a) low clouds, with a high solar albedo and high thermal emission temperature; and (b) high clouds, with a low solar albedo and low thermal emission temperature. The solar components are shown as straight arrows, and the infrared components, as curved arrows. The relative thicknesses of the arrows indicate the relative radiation intensities. The expected impact on surface temperature in each situation is noted along the bottom strip.

Clouds Radiative Forcing

$$\Delta R_{\mathrm{TOA}} = R_{\mathrm{Average}} - R_{\mathrm{Clear}}$$

$$R_{\mathrm{TOA}} = Q_{\mathrm{abs}} - \mathrm{OLR}$$

$$\Delta R_{\mathrm{TOA}} = \Delta Q_{\mathrm{abs}} - \Delta \mathrm{OLR}$$

- We can call the effect of clouds on the radiation budget the **cloud radiative effect (or forcing)** on the energy balance.
- The net cloud radiative effect is the difference between the net radiation at the **top of the atmosphere (TOA)** and what the net radiation would be if clouds were removed from the atmosphere leaving all else unchanged.
- It is possible to use Earth-orbiting satellites to accurately measure the radiative fluxes of energy entering and leaving Earth.
- The cloud-free scenes of the satellite observations can be averaged together to estimate the **clear-sky radiation budget**.
- If these cloud-free scenes are taken to represent the atmosphere in the absence of clouds, then the difference between the cloud-free radiation budget and the average of all scenes represents the effect of clouds on the radiation budget.

3.12 A SIMPLE MODEL FOR THE NET RADIATIVE EFFECT OF CLOUDINESS

(skip this Section)