Lecture two

Solar Radiation

2-1 Some Definitions:

Beam Radiation: The solar radiation received from the sun without scattered by the atmosphere. (Beam radiation is often referred to as *direct solar radiation*; to avoid confusion between subscripts for direct and diffuse, we use the term *beam radiation*.)

Diffuse Radiation: The solar radiation received from the sun after its direction has been changed by scattering in the atmosphere. (Diffuse radiation is referred to in some meteorological literature as ***sky radiation*** or ***solar sky radiation***; the definition used here will distinguish the diffuse solar radiation from infrared radiation emitted by the atmosphere.

Total Solar Radiation: The sum of the beam and the diffuse solar radiation on a surface. (The most common measurements of solar radiation are total radiation on a horizontal surface, often referred to as *global radiation on the surface*.)

Irradiance: The rate at which radiant energy is incident on a surface per unit Area of surface in units W/m^2. The symbol G is used for *solar irradiance*, with appropriate subscripts for ***beam***, ***diffuse***, or ***spectral radiation***. (G_b, G_d, G_λ)

Irradiation (J/m^2): The incident energy per unit area on a surface, found by integration of irradiance over a *specified time*, usually an hour or a day.

Insolation: term applying specifically to solar energy irradiation. The symbol **(H)** is used for insolation for a day. The symbol **(I)** is used for insolation for an hour (or other period if specified). The symbols **H** and **I** can represent *beam*, *diffuse*, or total and can be on surfaces of any orientation.

Radiosity or (Radiant Existence) (W/m^2): The rate at which radiant energy leaves a surface per unit area by combined *emission*, *reflection*, and *transmission*.

Emissive Power or (Radiant Self-Existence) (W/m^2): The rate at which radiant energy leaves a surface per unit area by ***emission only***.

Air Mass (m): Air mass is a measure of how far light travels through the Earth's atmosphere. The Sun is continually releasing an enormous amount of radiant energy into space. Earth receives a tiny fraction of this energy; yet, an average of 1367 watts

(W) reaches each square meter (m^2) of the outer edge of Earth's atmosphere. The atmosphere absorbs and reflects some of this radiation, including most X-rays and ultraviolet rays.

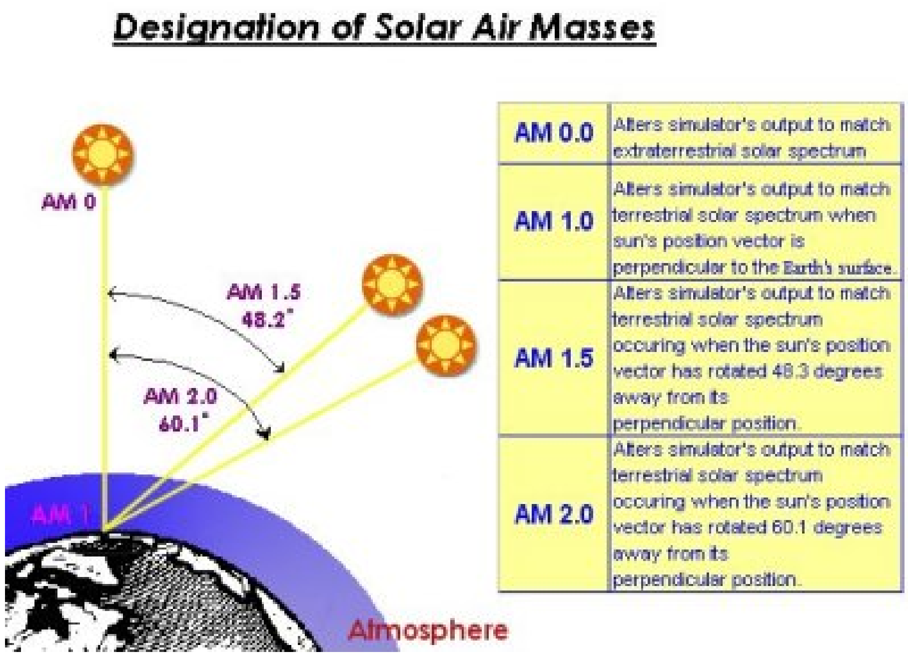


Figure 2.1: show change of air mass with zenith angle value

Q1) compare between direct solar radiation and sky radiation.

Q2) compare between irradiance and irradiation.

Q3) compare between irradiance and insolation.

Q4) compare between irradiance and radiosity.

Question: How much energy does light lose in traveling from the edge of the atmosphere to the surface of Earth?

This energy loss depends on the thickness of the atmosphere that the Sun's energy must pass through. The radiation that reaches sea level at high noon in a clear sky is 1000 W/m^2 and is described as "air mass 1" (or AM_1) radiation. As the Sun moves lower in the sky, the light passes through a greater thickness (or longer path) of air, losing more energy. For higher ***zenith angles***, the effect of the ***earth's curvature*** becomes significant and must be taken into account.

(Home work) Thus at sea level (m = 1) when the sun is at the zenith, and (m = 2) for a zenith angle θ_z of 60°. Air Mass = 1/cos (ZA°) where "ZA" stands for "zenith angle" which is how far away from directly overhead the sun.

Notes: $cos90 = 0$, $cos60 = \frac{1}{2}$, $\cos 30 = 0.877$, cos0= 1

2-2 Solar constant:

Radiation emitted from a spherical source decreases with increases square of the distance from the center of the sphere:

$$(Radiation\ emitted)\ E = E_1^*(\frac{R_1}{R_2})^2 \dots\dots\dots\dots\dots\dots\dots\dots. (2.1)$$

Where R is the radius from the center of the sphere (sun or earth orbit), and the subscripts denote two different distances from the center (R_1, R_2 refer to sun radius and Earth-orbital radius respectively). This is called the ***inverse square law***. From eq. (2.1) we expect that the radiative flux reaching the Earth's orbit is greatly reduced from that at the surface of the sun. The solar emissions of Fig. 2.2 must be reduced by a factor of **$2.167 x 10^{-5}$**, ***based on the square of the ratio of sun radius to Earth-orbital radius from eq. (2.1).*** This result is compared to the emission from Earth in Fig. 2.2.

$$\left(\frac{\textbf{sun radius}}{\textbf{earth orbit radius}}\right)^2 = (\frac{R_1}{R_2})^2 = \left(\frac{7 * 10^5}{1.5 * 10^8}\right)^2 = 21.777777777777777 * 10^{-6}$$

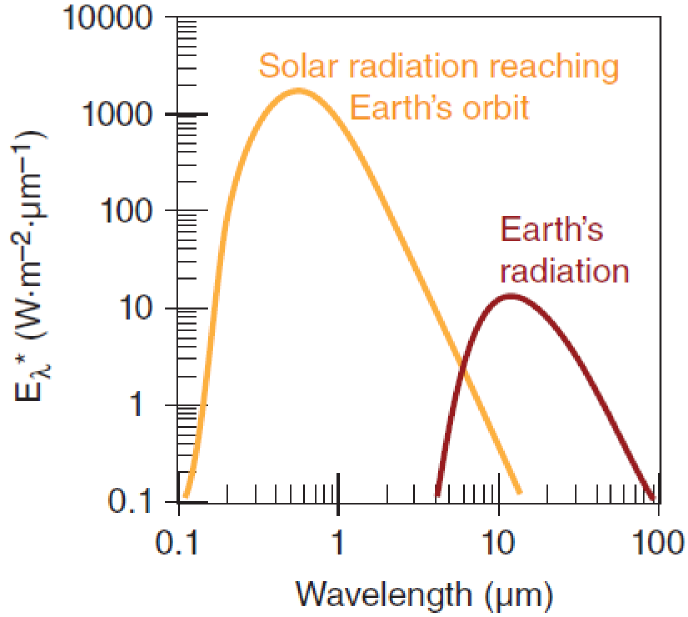


Figure 2.2

Blackbody radiance E reaching top of Earth's atmosphere from the sun and radiance of terrestrial radiation leaving the top of the atmosphere, plotted on a log-log graph.*

The area under solar-radiation curve in Fig. 2.2 is the total (all wavelengths) solar irradiance (TSI), So, reaching the Earth's orbit. We call it an **irradiance** here, instead of an **emittance**, because relative to the Earth it is an incoming radiant flux. This quantity was formerly called the ***solar constant*** but we now know that it varies slightly. The average value of solar irradiance measured at the Earth's orbit by satellites for a quiet sun (during sunspot minima) is about:

$$S_0 = 1361 \ (W.m^{-2}) \ \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots. (2.2)$$

In kinematic units (based on sea-level density), the solar irradiance is roughly So = 1.11 K·m s^{-1}. (**Home work**)

The solar constant G_{sc} is the energy from the sun per unit time received on a unit area of surface perpendicular to the direction of propagation of the radiation at mean earth-sun distance outside the atmosphere.

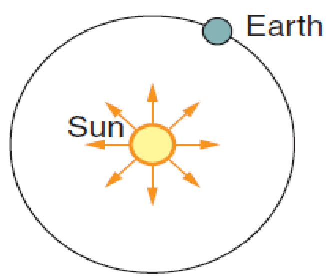
The availability of very high altitude aircraft, balloons, and spacecraft has permitted direct measurements of ***solar radiation outside*** most or all of the earth's atmosphere. They resulted in a value of the solar constant G_{sc} of **1361 W/m²** with an estimated error of ±1.5%.

Problem 1:

Estimate the value of the solar irradiance reaching the orbit of the Earth, given a sun surface temperature (5770 K), sun radius (6.96x10^{5}km), and orbital radius (1.495x10^8 km) of the Earth from the sun.

The answer:

Given: T_{sun} = 5770 K
R_{sun} = 6.96x10^5 km = solar radius
R_{Earth} = 1.495x10^8 km = Earth orbit radius
Find: S_o = ? W·m^{-2}
Sketch:



By using equ: $E^* = \sigma_{SB} \cdot T^4$

$$E_1^* = (5.67 * 10^{-8}\ W.m^{-2}K^{-4})(5770K)^4 = 6.285 * 10^7\ W.m^{-2}$$

By assumed $R_1 = R_{sum}$ and $R_2 = R_{earth}$

$$S_0 = E_2^* = (6.285 * 10^7\ W.m^{-2}) * \left(\frac{6.96 * 10^5 km}{1.495 * 10^8\ km}\right)^2 = \mathbf{1362}\left(\frac{\boldsymbol{watt}}{\boldsymbol{m^2}}\right)$$

If this radiation strikes a surface that is not perpendicular to the radiation, then the radiation per unit surface area is reduced according to the **sine law**. The resulting flux, F_{rad}, at this surface is:

$$F_{rad.} = E.\sin(\Psi) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots..(2.3)$$

Where Ψ is the *elevation angle* (the angle of the sun above the surface).

Problem 2:

During the equinox at noon at latitude ϕ =60°, the solar elevation angle is Ψ = 90° – 60°= 30°. If the atmosphere is perfectly transparent, then how much radiative flux is absorbed into a perfectly black asphalt parking lot?

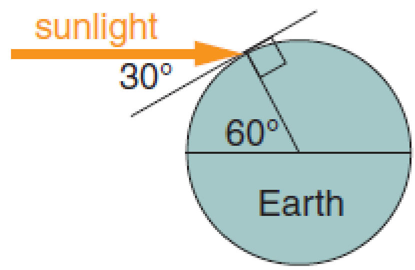
Answer

Given: Ψ = 30° = elevation angle

$E = S_o = 1361\ W\cdot m^{-2}$. Solar irradiance

Find: $F_{rad} = ?\ W\cdot m^{-2}$

$$F_{rad} = (1361\ W\cdot m^{-2})\cdot\sin(30°) = \underline{680.5\ W\cdot m^{-2}}.$$



2.3 Insolation:

- It is a quantity amount of incident solar power on a unit surface commonly expressed in units of kw/m².
- Due to atmospheric effect, the peak solar insolation incident on a terrestrial surface oriented to the ***sun at noon*** on a ***clear day*** is on order of **1000W/m²**.
- A solar insolation level of 1kw/m² is called ***peak sun solar insolation*** is denoted by (I).
- The graph show given the amount of power present in different wavelengths of radiation.
- It can be seen from the fig 2.3, that 50% solar energy is in the form of *thermal energy*.

- Solar PV captures the energy invisible region, solar thermal capture energy in infrared region.

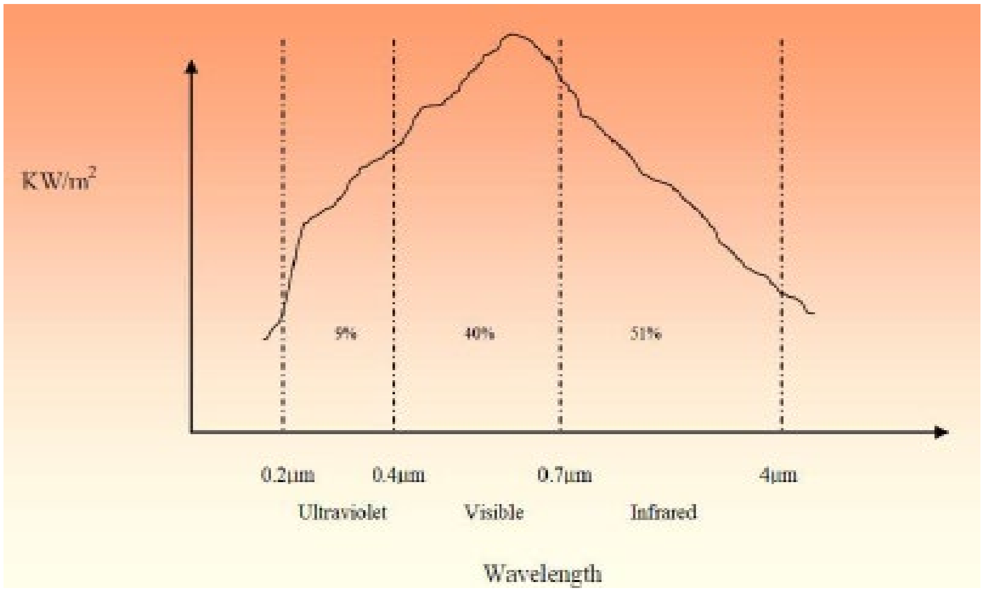


Figure 2.3: Distribution of solar power

2.4 Average Daily Insolation:

The average daily insolation $\bar{E}$ depended basically on both the solar elevation angle (which varies with season and time of day) and the duration of daylight. For example, there is more total insolation at the poles in summer than at the equator, But the low sun angle near the poles is more than (small) and compensated by the long periods of daylight. The equation to calculate insoluation angle can be given as:

$$\bar{E} = \frac{S_o}{\pi} \cdot \left(\frac{a}{R}\right)^2 \cdot \left[h_o{}' \cdot \sin(\phi) \cdot \sin(\delta_s) + \cos(\phi) \cdot \cos(\delta_s) \cdot \sin(h_o) \right] \qquad (2.21)$$

Where S_o =1361 W m^{-2} is the solar irradiance, a = 149.457 Gm is Earth's semi-major axis length, R is the actual distance for any day of the year, ϕ = latitude.

$$R = a \cdot \frac{1-e^2}{1+e \cdot \cos(\nu)}$$

where e = 0.0167 is eccentricity, and a = 149.457 Gm is the semi-major axis length. If the simple approximation of $\nu \approx M$ is used, then angle errors are less than 2° and distance errors are less than 0.06%.

In, h'_0: is the sunset and sunrise hour angle in radians. The hour angle h_o at sunrise and sunset can be found using the following steps:

$\boldsymbol{\alpha = -\tan(\phi).\tan(\delta_s)}$

$\boldsymbol{h_0 = \arccos(\alpha)}$

Fig. 2.4 shows the average incoming solar radiation vs. ***latitude and day*** of the year, found using eq. (2.21). For any one hemisphere, *E* has greater difference between equator and pole during winter than during summer. This causes stronger winds and more active extratropical cyclones in the winter hemisphere more active, than in the summer hemisphere.

[CAUTION]. When finding the **arccos**, your answer might be in degrees or radians, depending on your calculator. In all cases if necessary, convert the hour angle to units of radians, the result of which is ho'.]

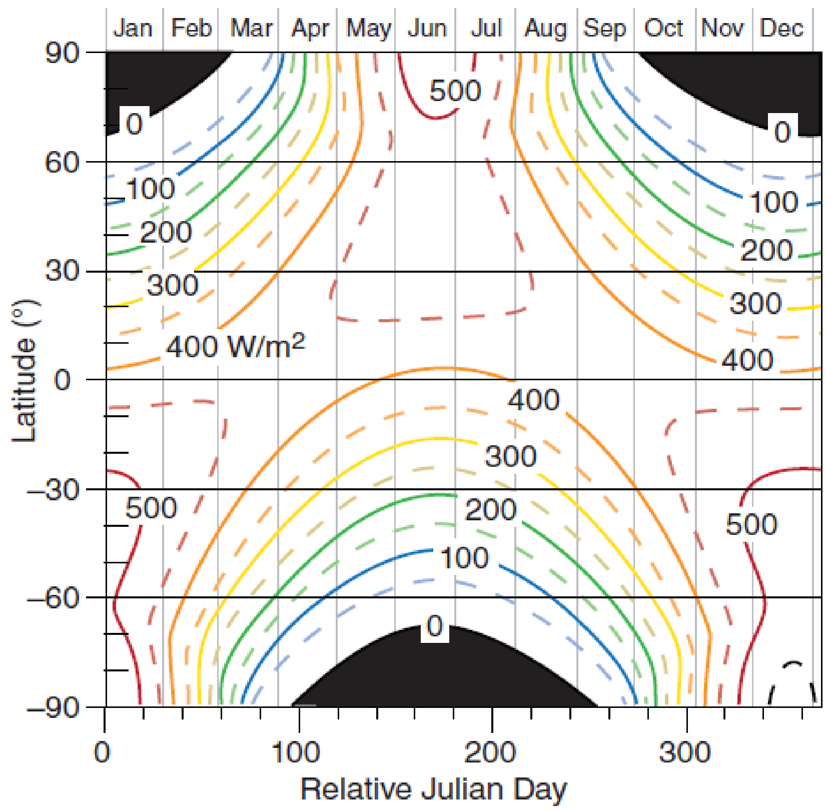


Figure 2.11
Average daily insolation $\overline{E}$ (W m^{-2}) over the globe.

Figure 2.4: show the calculated values of average daily insolation

Problem:

if the sun radius is $7.2*10^{5}$ km, the solar radiation emitted from the sun surface is $5.7*10^{7}$watt/m^2, what is the solar radiation incident on the earth surface at noon and equinox , if the solar elevation is 60 degree and atmosphere is perfectly transparent.

Problems

Find the average daily insolation over Vancouver during the summer solstice.

Find the Answer:

Given: $d = d_r = 173$ at the solstice,
$\phi = 49.25°N$, $\lambda_e = 123.1°W$ for Vancouver.
Find: $\overline{E} = ?\ W\ m^{-2}$

Use eq. (2.5): $\delta_S = \Phi_r = 23.45°$
Use eq. (2.2): $M = 167.55°$, and assume $\nu \approx M$.
Use eq. (2.4): $R = 151.892$ Gm.
Use eq. (2.22):
$h_o = \arccos[-\tan(49.25°)\cdot\tan(23.45°)] = 120.23°$
$h_o' = h_o \cdot 2\pi/360° = 2.098$ radians
Use eq. (2.21):

$$\overline{E} = \frac{(1361 W\cdot m^{-2})}{\pi}\cdot\left(\frac{149 Gm}{151.892 Gm}\right)^2\cdot [2.098\cdot\sin(49.25°)\cdot\sin(23.45°) + \cos(49.25°)\cdot\cos(23.45°)\cdot\sin(120.23°)]$$

$$\overline{E} = (1320\ W\ m^{-2})\cdot[2.098(0.3016)+0.5174]$$

$$= \underline{483.5\ W\ m^{-2}}$$

2.4 Vertical Profile of Absorption of trace gases (lower atmosphere):

The atmosphere interacts with both incoming solar radiation and outgoing terrestrial radiation. The strength of the interaction as a function of wavelength responsible for heating of the lower atmosphere.

2.4.1 Shortwave Radiation

Blackbody curves for emitters at 5777 K and 280 K are shown in figure 3, above it Also shown is the fraction of light entering the top of the earth's atmosphere that is absorbed before reaching 10 km (middle panel) and sea level.

At 10 km (top of the troposphere) virtually all radiation below **290 nm** has been absorbed. All radiation below **100 nm** is absorbed in the thermosphere above **100 km.**

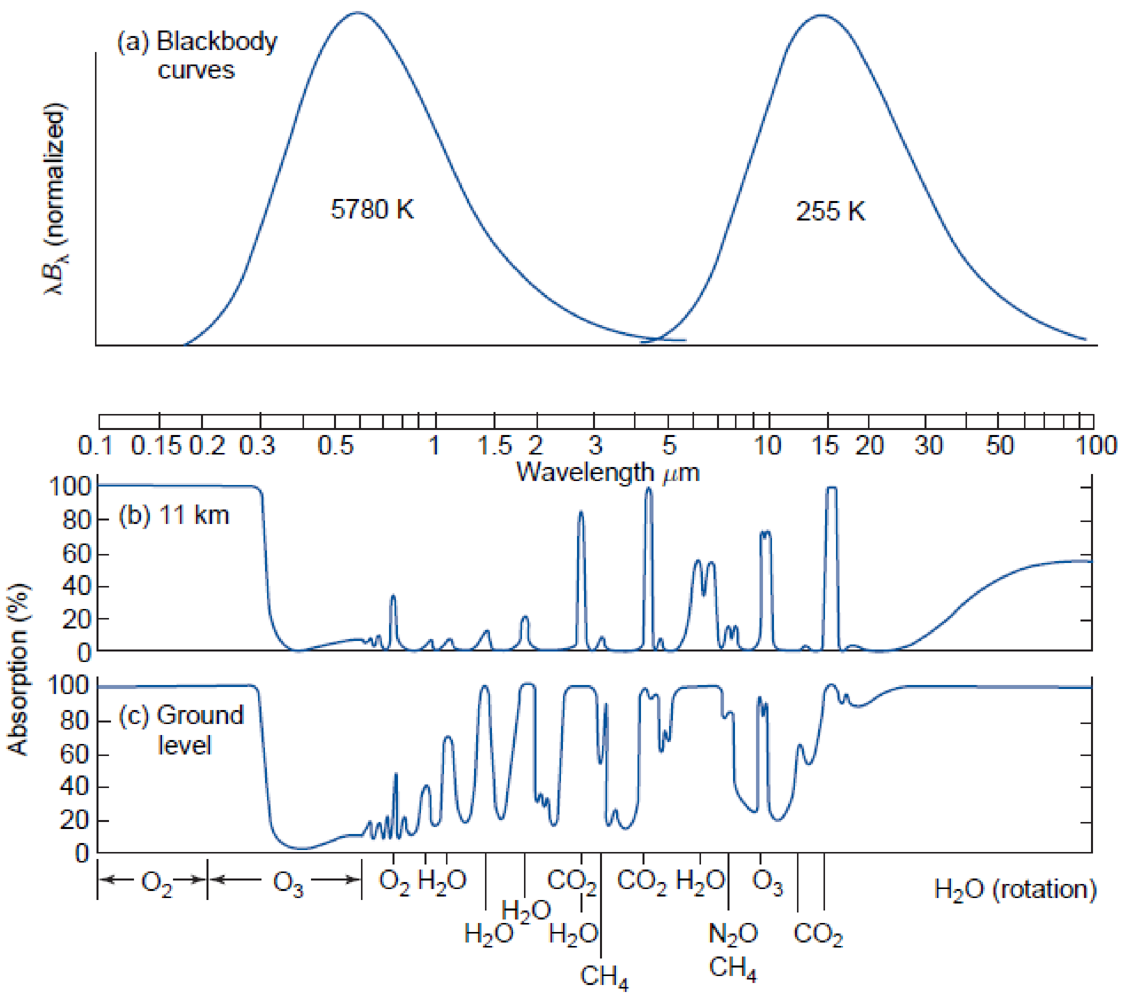


Figure 3: Absorptivity of solar radiation according to wavelength

There is little tropospheric absorption below 0.6um or 600 nm ($1m=10^9nm$) but H_2O and CO_2, at high tropospheric concentrations, deplete the (near IR part) of the incoming solar flux appreciably.

2.4.2 Long wave Radiation (outgoing radiation or Terrestrial Radiation):

1. Much of the outgoing radiation is absorbed in the lowest 10 km where several different molecules are efficient absorbers of upwelling IR radiation much of the outgoing radiation of wavelengths less than ***7 µm*** is absorbed by **water vapor**, with some contribution from ***methane and nitrous oxide, N_2O***.

2. Light of wavelengths longer than **13 µm** is efficiently absorbed by **CO_2**. This band, center at **15 µm**, is important as it lies close to the maximum of the longwave irradiance spectrum. At longer wavelengths **water vapour** is excited into many rotational states that effectively form an absorption continuum beyond **25 µm**.

3. The only fraction of the outgoing radiation that is transmitted through the troposphere without undergoing appreciable absorption lies in the so-called **atmospheric window between 7 and 13 µm.**

The only significant absorptions of infra-red radiation in the stratosphere are due to **ozone**. The **9.6 µm band** of ozone happens to lie in the middle of the atmospheric window and as a result means that stratospheric ozone plays a significant role in the outgoing longwave radiation budget of the Earth.

2.3. Absorption by Stratospheric (upper Temperature atmosphere)

1. The main layer of ozone in the atmosphere is situated between 15 and 30 km and reaches a maximum concentration of around ***5 x 10^{12} molecules cm^{-3}*** at 22 km.
2. Zone is a very efficient absorber of solar radiation between ***200 and 300 nm***.
3. The maximum temperature at the top of the stratosphere occurs at around 50 km, well above the main *ozone layer*.
4. To understand the effect of stratospheric ozone on the temperature profile we need to understand the way ozone is created and destroyed in the mesosphere and stratosphere.
   The absorption of UV radiation by both oxygen and ozone leads to their photolysis and the energy involved in these sunlight-induced reactions produces local warming. *The temperature at a particular altitude will then be a*

combination of the rates of photolysis of the two oxygen species, in particular ozone, and the air density. The rates of photolysis will depend on the local incidence of radiation and thus on the optical density of the atmosphere in the column above at a given wavelength. Although the temperature profile is strongly linked to that of ozone its maximum occurs not at the maximum ozone concentration, but above it and close to the region where the photolytic formation and loss processes of ozone are most.

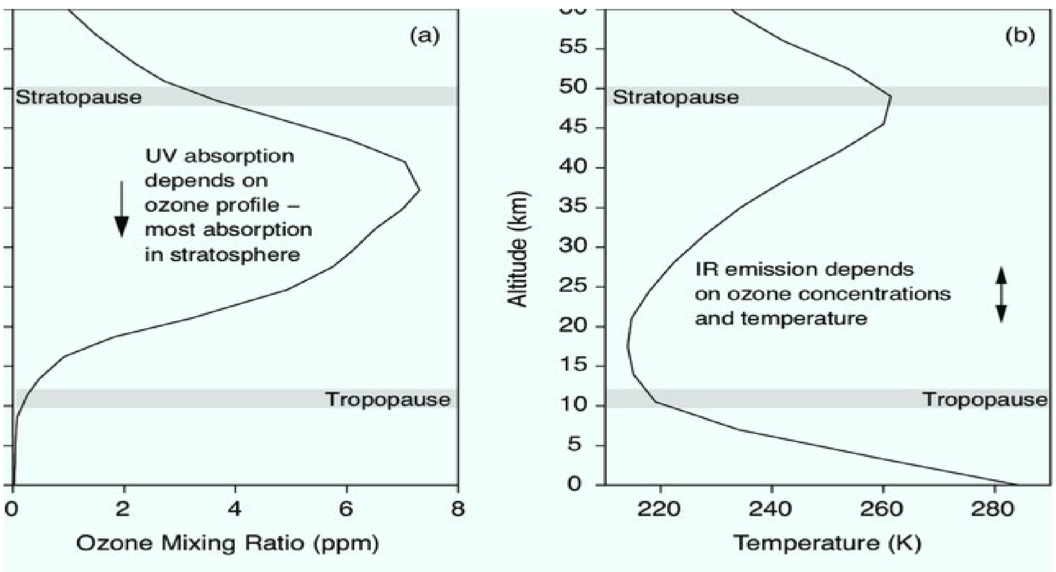


Figure 4: Vertical profiles of ozone-related quantities. (a) Typical mid-latitude ozone mixing ratio profile, (b) atmospheric temperature profile, based on Fleming et al. (1990), showing the stratosphere bounded by the tropopause below and the stratopause above;

2.4 Solar Heating Rates (Absorption and Emission of Infrared Radiation)

The two previous subsections dealt with the scattering and absorption of radiation in planetary atmospheres in the absence of **emission**. The relationships discussed in those subsections are applicable to the transfer of solar radiation through planetary atmospheres. The remains of this section is concerned with the absorption and emission of infrared radiation in the absence of scattering. This simplified treatment is justified by the fact that the wavelength of infrared radiation is very long in comparison to the circumference of the air molecules so the scattering efficiency is negligible.

Q) Scattering efficiency is negligible in infrared radiation?

First we will derive the equation that governs the transfer of infrared radiation through a gaseous medium.
Rewriting eq.
$[dI_\lambda = -I_\lambda \rho r k_\lambda ds]$
the rate of change of the monochromatic intensity of *outgoing terrestrial radiation* along the path length *ds* due to the absorption within the layer is:

$$dI_\lambda(absorption) = -I_\lambda \rho r k_\lambda ds = -I_\lambda \alpha_\lambda \qquad (2.1)$$

[Where ρ is the density of the air, r is the mass of the absorbing gas per unit mass of air, and k_λ is the mass absorption coefficient, which has units of $m^2\ kg^{-1}$]
Where α_λ is the absorptivity of the layer.

the corresponding rate of change due to the emission of radiation is:

$$dI_\lambda(emission) = B_\lambda(T)\varepsilon_\lambda \qquad (2.2)$$

$B_\lambda(T)$: Emission function

Appealing Kirchhoff's law $[\alpha_\lambda = \varepsilon_\lambda]$ and summing the two expressions, we obtain Schwarzschild's equation:

$$dI_\lambda = -\left(I_\lambda - B_\lambda(T)\right) k_\lambda \rho r ds \qquad (drive)\,?\,(h.w) \qquad (2.3)$$

We can put some notes about this equations such as:

1- Optical depth is a measure of how much light is absorbed or scattered as it passes through a medium, such as a gas or a solid material. It is a dimensionless quantity that is defined as the negative natural logarithm of the fraction of incident light that makes it through the material.

2- Mathematically, the optical depth τ is given by:

$$\tau = -\ln(I/I_0)$$

Where I_0 is the intensity of the incident light and I is the intensity of the light after it has passed through the medium. The optical depth depends on the properties of the medium, such as its density, composition, and thickness, as well as the wavelength of the light.

3- A high optical depth means that a large fraction of the light is absorbed or scattered, while a low optical depth means that most of the light passes through the material unaffected. Optical depth is an important concept in astronomy, atmospheric science, and other fields where the interaction of light with matter is of interest.

4- In absorbed solar radiation, optical depth is strongly dependent on zenith angle. This dependence is exploited in the design of Satellite area *scanning* radiometers, which monitor radiation emitted along very long, oblique path lengths through the atmosphere.

5- Optical depth for emitted radiation is strongly dependent upon the *wavelength of the radiation*.

6- Centers of absorption lines encountered much higher in the atmosphere density than in the gaps between the lines.
7- Within these so-called Atmospheric *windows of the electromagnetic spectrum*, an appreciable fraction of the radiation emitted from the Earth's surface escapes directly to space without absorption during its passage through the atmosphere.
8- Optical depth dependent upon the vertical profiles of the concentrations r of the various greenhouse gases.

b. The plane-parallel

Many atmospheric radiative transfer calculations can be simplified by use *plane-parallel approximation* in which temperature and the densities of the various atmospheric constituents are assumed to be functions of height (or pressure) only. With this simplification, the flux density passing through a given atmospheric level is given by:

$$F_v^{\downarrow\uparrow}(\tau_v) = \int_{2\pi} I_v{}^{\downarrow\uparrow}(\tau_v, cos\theta) cos\theta d\omega \qquad (2.5)$$

Where τ_v, the normal optical depth as defined in eq. $\tau_\lambda = \int_z^\infty k_\lambda \rho r dz$, is used as a vertical coordinate, and the arrows (↓↑) denote that both upward and downward fluxes are taken into account, $d\omega$ represents an elemental arc of solid angle. Since this section focuses on infrared radiative transfer, we use wave number ($v = 1/\lambda$) notation to be consistent with the literature in this subfield. Integrating over azimuth angle and denoting $\mathbf{cos\theta}$ by $\boldsymbol{\mu}$, we obtain:

$$F_v^{\downarrow\uparrow}(\tau_v) = 2\pi \int_0^1 I_v{}^{\downarrow\uparrow}(\tau_v, \mu)\mu d\mu \qquad (drive\,?) \qquad (2.6)$$

The monochromatic intensity in this expression may be broken down into three components:

• Upward emission from the Earth's surface that reaches level without being absorbed.
• Upward emission from the underlying atmospheric layer.
• Downward emission from the overlying layer.

The expressions used in evaluating these terms are analogous to Beer's law, with the intensity transmissivity T_v replaced by the flux transmissivity:

$$T_v^f = 2\int_0^1 e^{-\tau_v/\mu}\,\mu d\mu \qquad (2.7)$$

For many purposes it is sufficient to estimate the flux transmissivity from the approximate formula

$$T_v^f = e^{-\tau_v/\bar{\mu}} \qquad (2.8)$$

Where the "average" or "effective zenith angle" is:

$$\frac{1}{\bar{\mu}} = sec53^0 = 1.66 \qquad (2.9)$$

In other words, the flux transmissivity of a layer is equivalent to the intensity transmissivity of parallel beam radiation passing through it with a zenith angle 53°. The factor $\frac{1}{\bar{\mu}}$ is widely used in radiative transfer calculations and is referred to as the diffusivity factor. Applying the concept of a diffusivity factor (2.6) can be integrated over to obtain:

$$F_v^{\downarrow\uparrow}(\tau_v) = \pi I_v^{\downarrow\uparrow}(\tau_v, \bar{\mu}) \qquad (2.10)$$

Problems

1- Compare between beam radiation and diffuse radiation.

2.5 Vertical Profiles of Radiative Heating Rate: (تكملة فيما بعد)

The radiatively induced time rate of change of temperature due to the absorption or emission of radiation within an atmospheric layer is given by:

$$\rho c_p \frac{dT}{dt} = -\frac{dF(z)}{dz} \qquad (2.11)$$

$$\left(\frac{dT}{dt}\right)_v = -\frac{1}{\rho c_p}\frac{dF_v(z)}{dz} = -\frac{1}{\rho c_p}\frac{d}{dz}\left[\int_{4\pi} I\,\mu\,d\omega\right] =$$

$$\left(\frac{dT}{dt}\right)_\nu = -\frac{1}{\rho c_p}\frac{dF_\nu(z)}{dz}$$

$$= -\frac{1}{\rho c_p}\frac{d}{dz}\left[\int_{4\pi} I_\nu \mu\, d\omega\right]$$

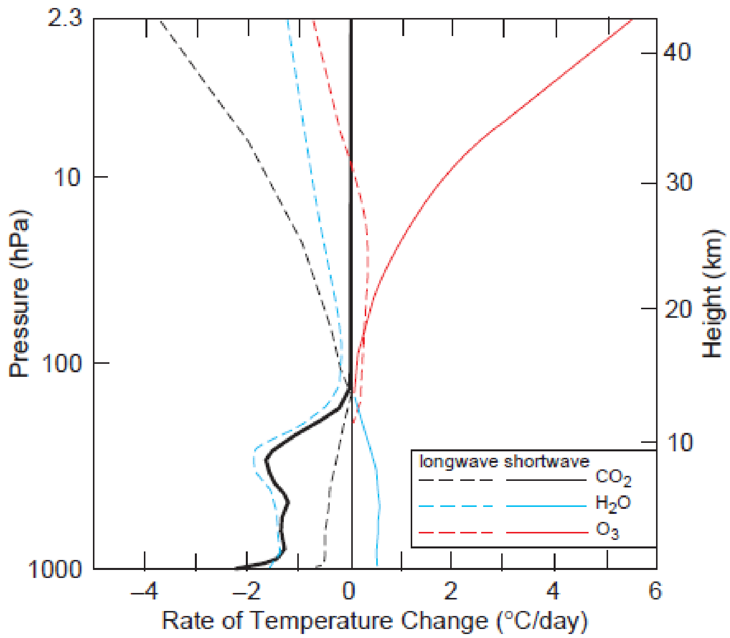
$$= -\frac{1}{\rho c_p}\int_{4\pi}\frac{dI_\nu}{ds}d\omega$$

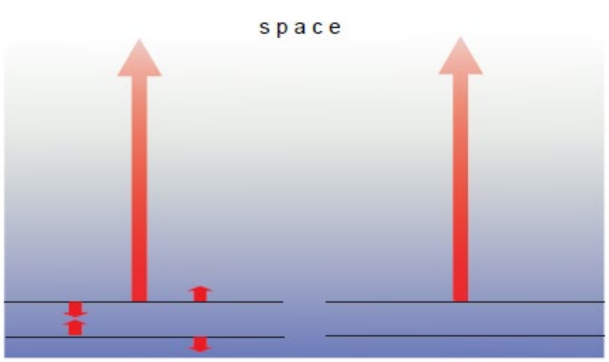
$$= -\frac{2\pi}{\rho c_p}\int_{-1}^{1}\frac{dI_\nu}{ds}d\mu$$

$$\left(\frac{dT}{dt}\right)_\nu = \frac{2\pi}{c_p}\int_{-1}^{1} k_\nu r\,(I_\nu - B_\nu)\,d\mu$$

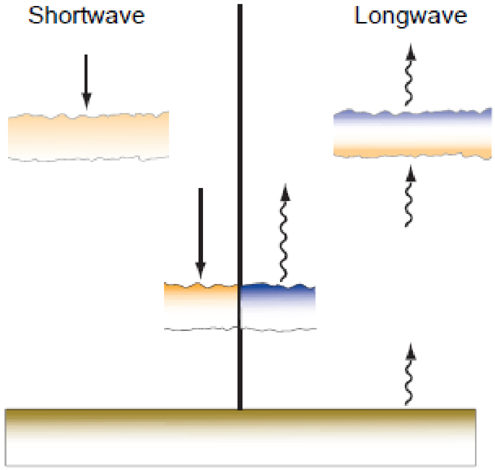
$$\left(\frac{dT}{dt}\right)_\nu = -\frac{2\pi}{c_p}\int_{0}^{1} k_\nu r B_\nu(z) e^{-\tau_\nu/\mu}\,d\mu$$

$$\left(\frac{dT}{dt}\right)_\nu = -\frac{\pi}{c_p} k_\nu r B_\nu(z)\frac{e^{-\tau_\nu/\overline{\mu}}}{\overline{\mu}}$$





(a) (b)


Shortwave
Longwave