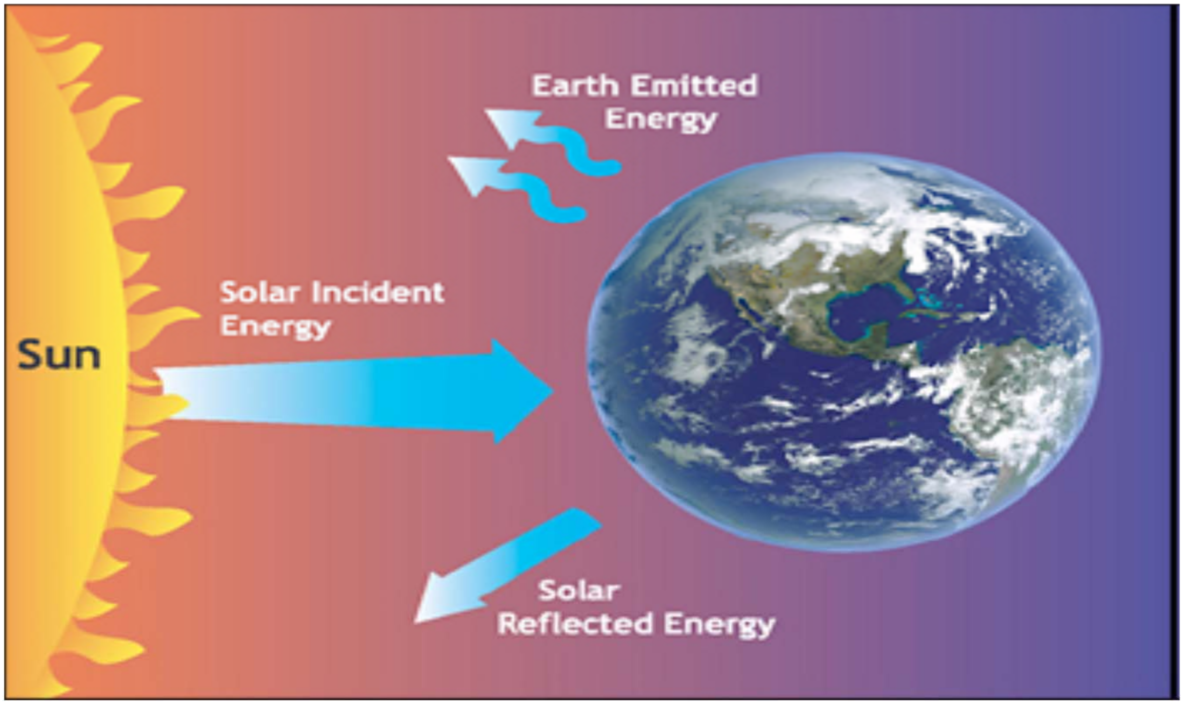
AOSC400-2015
September 29, Lecture # 8

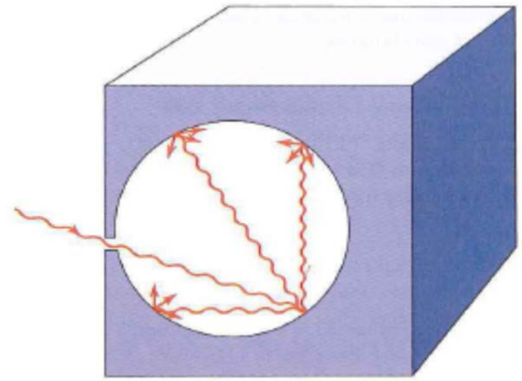
- Radiative Properties of Non-black Materials
- Simple radiation balance climate model with a greenhouse effect
- Physics of Scattering and Absorption and Emission
- Simple aspects of radiative transfer in the visible
- Beer-Bouguer-Lambert Law
- Langley Plots

What drives climate to change? - **“radiative forcing”**: A small change in the energy balance of the earth can change surface temperatures, winds, precipitation patterns => **it’s climate**



Subsequent lectures lead to improved understanding of the concept of “Radiative Forcing”.

Radiative Properties of Nonblack Materials



- Blackbodies absorb all incident radiation
- Nonblack bodies such (e.g., gaseous media) can also reflect and transmit radiation.
- Their behavior can be represented by applying the radiation laws derived for blackbodies. For this purpose it is useful to define:
- the (monochromatic) *emissivity* ε_λ, i.e., the ratio of the monochromatic intensity of the radiation emitted by the body to the corresponding blackbody radiation:

$$\varepsilon_\lambda = I\lambda\,(emitted)/B\,\lambda\,(T) \qquad (4.13)$$

Then: $F_E = \varepsilon_\lambda \sigma T^4$

or:

$$E_\lambda(T) = \epsilon_\lambda(T) B_\lambda(T),$$

For infrared radiation, the radiative behavior of fresh snow is similar to a blackbody.

Surface	ε_{IR}	Surface	ε_{IR}
Sea	0.95–1	Grass	0.90–0.95
Fresh snow	0.99	Desert	0.85–0.90
"Old" snow	0.80	Forest	0.95
Liquid water clouds	0.25–1	Concrete	0.70–0.90
Cirrus	0.10–0.90	Urban	0.85

Typical values of the emissivity in the IR region for several surface types.

The (monochromatic) *absorptivity, reflectivity, and transmissivity, namely,* the fractions of the incident monochromatic intensity that a body absorbs, reflects, and transmits, are:

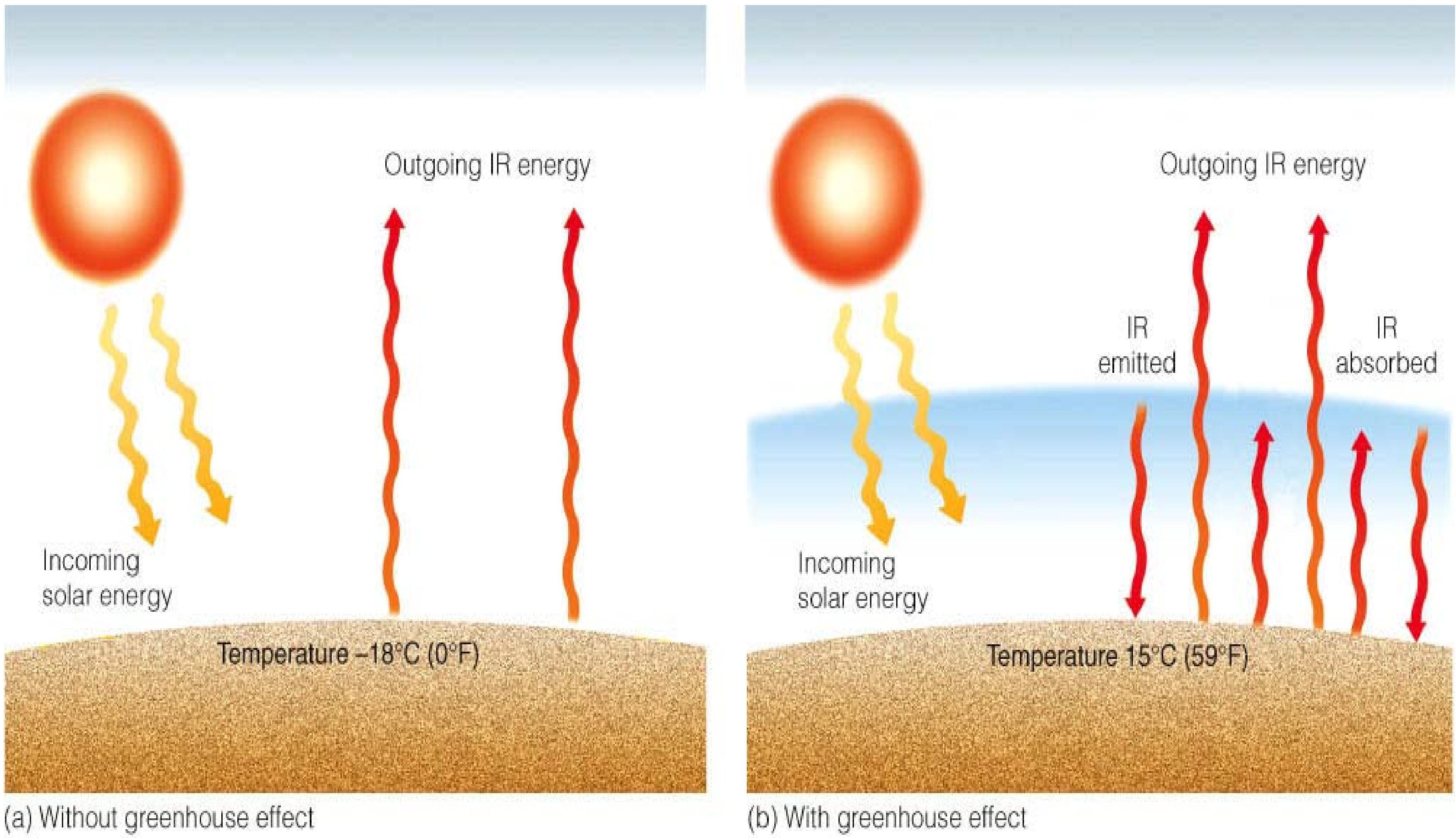
$$\alpha_\lambda = \frac{I_\lambda(\text{absorbed})}{I_\lambda(\text{incident})}, \quad R_\lambda = \frac{I_\lambda(\text{reflected})}{I_\lambda(\text{incident})}$$

$$\text{and} \quad T_\lambda = \frac{I_\lambda(\text{transmitted})}{I_\lambda(\text{incident})} \qquad (4.14)$$

The Greenhouse Effect

- Water vapor, carbon dioxide, and other gases absorb radiation more strongly in the longwave part of the spectrum than the shortwave part.
- Incoming solar radiation passes through the atmosphere quite freely, whereas terrestrial radiation emitted from the Earth's surface is absorbed and re-emitted in its upward passage through the atmosphere.
- Following shows how such greenhouse gases in the atmosphere tend to warm the surface of the planet.

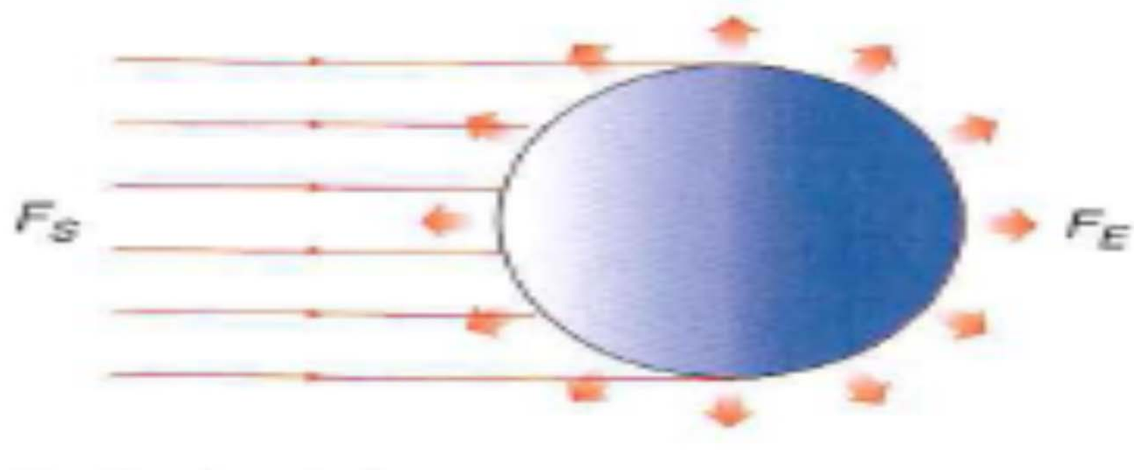
"The Greenhouse Effect"



$T_e = [(1-r)S/4\sigma]^{1/4}$; **Actual value of T_e is 15^0 C.**

Important to note:

The solar radiation is intercepted over an area of πR^2_E but the outgoing terrestrial (longwave) is emitted from the entire Earth from $4\pi R^2_E$, shown in detail in next slide:



$$F_E = \sigma T_E^4 = \frac{(1 - A)F_s}{4} = \frac{(1 - 0.30) \times 1368}{4}$$

$$= 239.4 \text{ W m}^{-2}$$

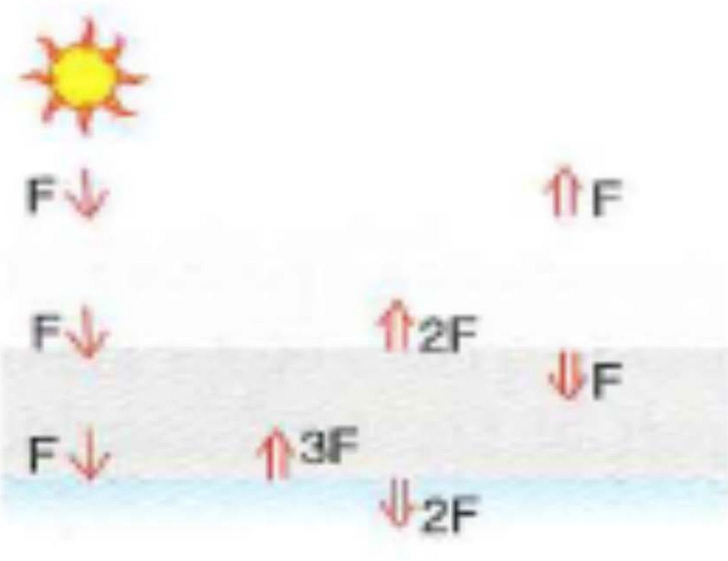
Solving for T_E, we obtain

$$T_E = \sqrt[4]{\frac{F_E}{\sigma}} = \left(\frac{239.4}{5.67 \times 10^{-8}}\right)^{1/4} = 255 \text{ K}$$

Sometimes, in terms of the black-body concept, the total emitted from black-body is written as:

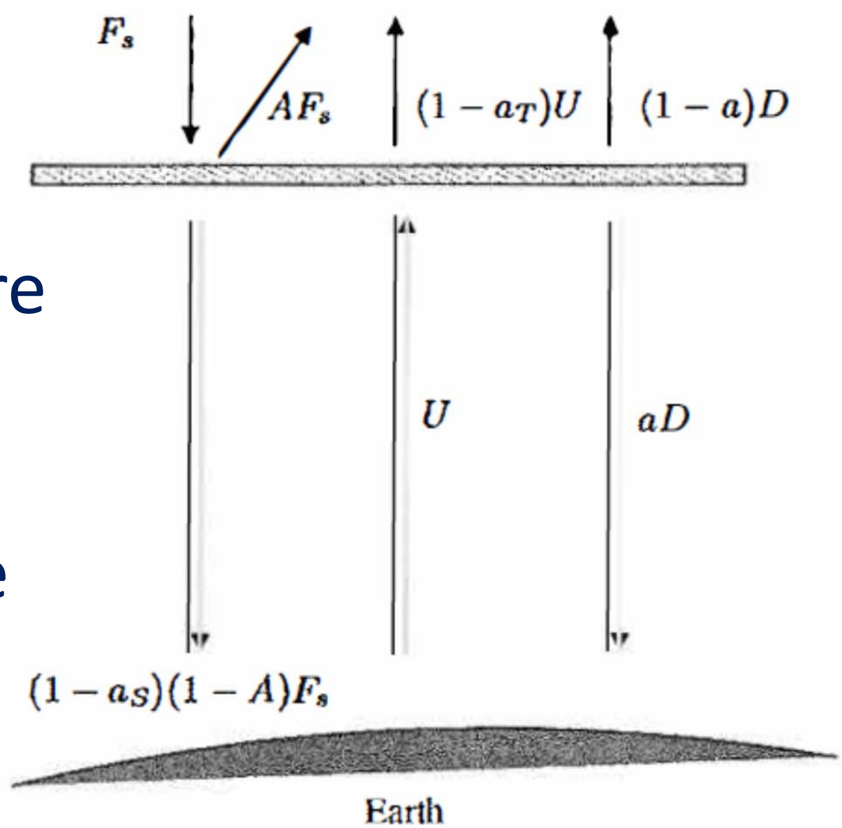
$$B(T) = \int_0^{+\infty} B_\lambda(T)\,\mathrm{d}\lambda = \sigma T^4.$$

Assume that the atmospheric layers and the surface of the planet are in radiative equilibrium. How is the surface temperature of the planet affected by the presence of this atmosphere?



Replace the textbook discussion of above by following:

Consider the atmosphere as a layer at a given distance from the Earth's surface. Let us assume that nothing happens in the region between the surface and the layer . From the radiative properties of the atmosphere (strong absorption of the infrared radiations and atmospheric window for the visible radiations), we assume that:



From Chapter 2, B. Sportisse.

- The layer and the Earth are at radiative equilibrium
- The layer reflects a fraction A of the incident solar radiation Fs. It absorbs a fraction a_s and transmits to the Earth a fraction $(1 - a_s)$ of the remaining radiation $(1 - A)\ F_{s.}$ The Earth is supposed to absorb the whole received radiation;
- the Earth emits longwave radiations U (up): a fraction (a_T) is absorbed by the layer while the remaining part ($1 - a_T$) is transmitted to space;
- the layer is then heated and emits a radiation D (down): a fraction a is transmitted to the Earth and then absorbed.

Let us investigate the sensitivity of the Earth's temperature (T) with respect to the absorption coefficient of the layer for the terrestrial radiation (a_T).

Taking
data from this figure:

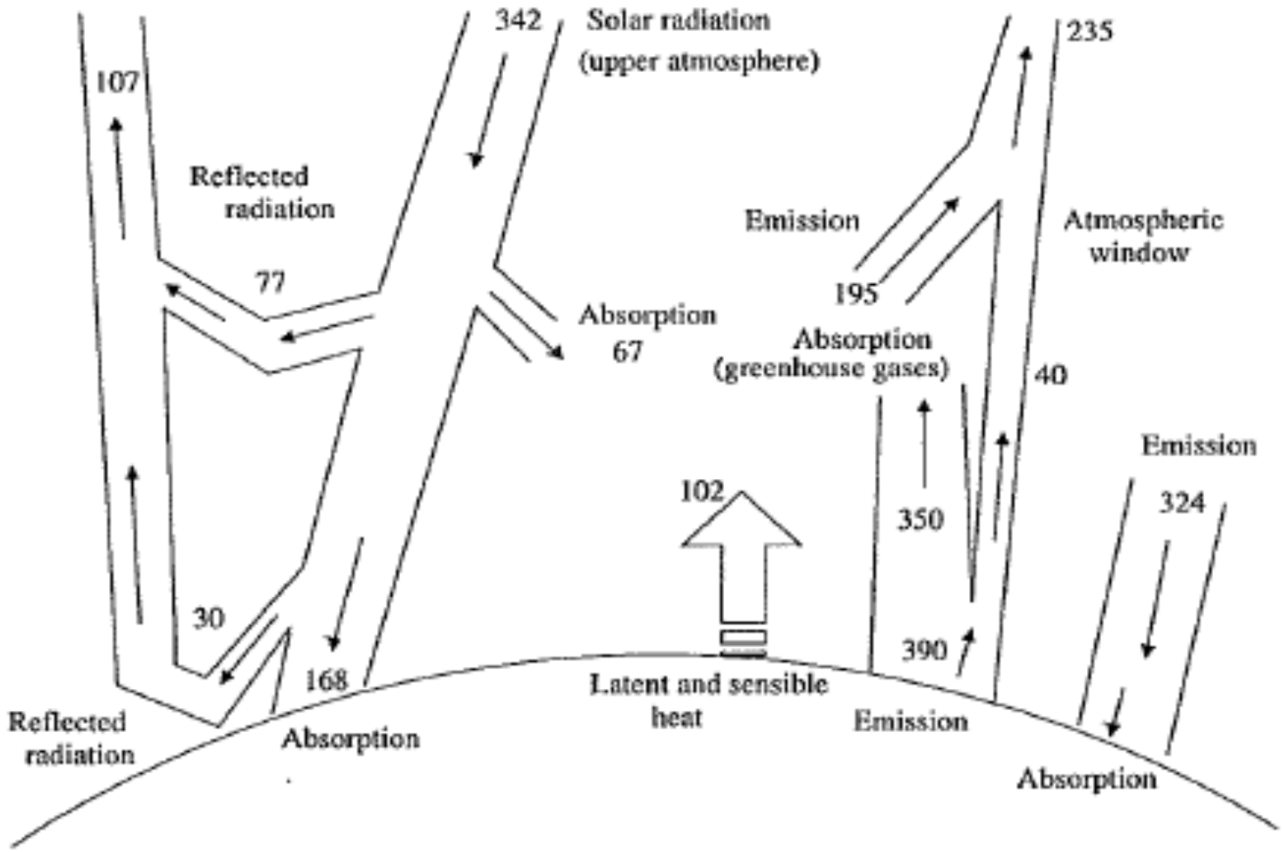


Fig. 2.14 Global energy budget for the Earth/atmosphere system. The fluxes are expressed in W m^{-2}. The values are indicative. Source: [74]

We get:

$$a_T = \frac{350}{390}, \qquad a_S = \frac{198}{265}, \qquad a = \frac{324}{519}.$$

The budget energy for the Earth and the atmospheric layer reads

$$\begin{cases} (1 - a_S)(1 - A)F_s + aD = U = \sigma T^4 \\ a_S(1 - A)F_s + a_T U = D. \end{cases}$$

Thus,

$$\sigma T^4 = U = \frac{(1 - a_S) + a\, a_S}{1 - a\, a_T}(1 - A)\, F_s. \tag{2.46}$$

The function $T(a_T)$ is an *increasing* function with respect to the absorption coefficient of the layer (a_T). The more absorbing the layer is, the higher the Earth's temperature is: this is the *greenhouse effect*, to be measured by the difference between the emission effective temperature T_e given by (2.42) and the surface temperature T. This gives straightforward

$$T = \left(\frac{(1 - a_S) + a\, a_S}{1 - a\, a_T}\right)^{0.25} T_e. \tag{2.47}$$

As $T_e \simeq 255\,\mathrm{K}$, we get $T \simeq 288\,\mathrm{K}$ (not far from the observed value).

Physics of Scattering and Absorption and Emission

- The scattering and absorption of radiation by gas molecules and aerosols all contribute to the extinction of the solar and terrestrial radiation passing through the atmosphere.
- Each of these contributions is linearly proportional to:

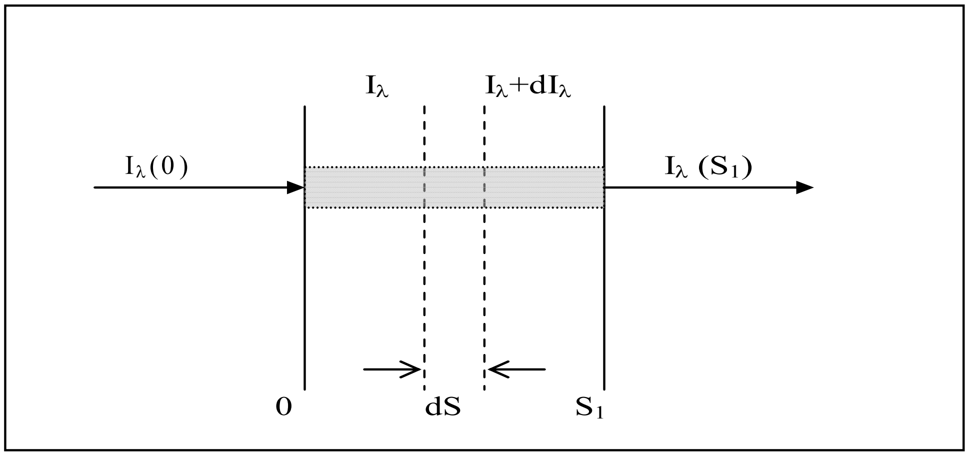
(1) the intensity of the radiation at that point along the ray path

(2) the local concentration of the gases and/or particles that are responsible for the absorption and scattering, and:

(3) the effectiveness of the absorbers or scatterers.

Simple Aspects of Radiative Transfer

Tracing a ray from the sun ($I_\lambda(0)$) through a medium:



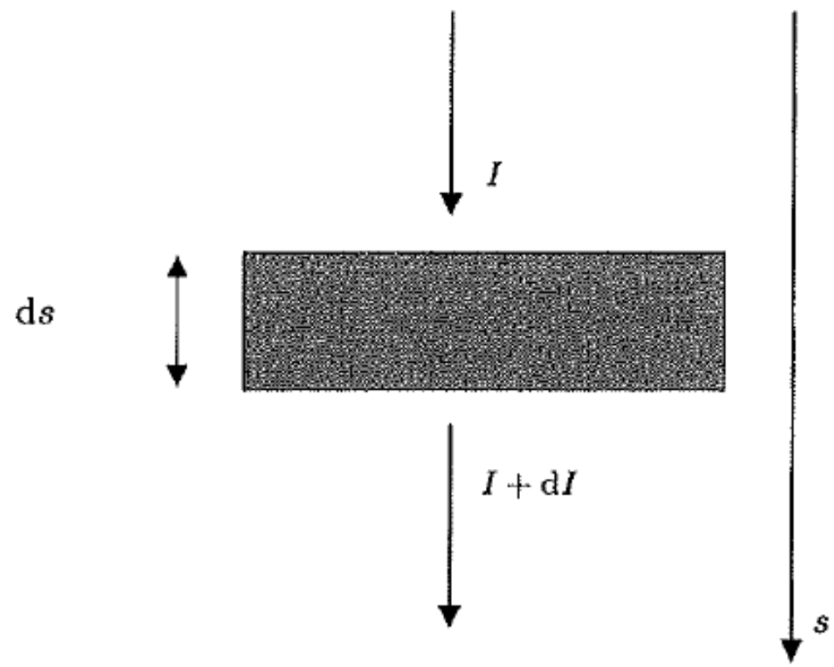
The equation of transfer

Depletion of energy takes place when the light beam goes through the medium:

$$I_\lambda \rightarrow I_\lambda + dI_\lambda$$

$$dI_\lambda \propto \rho I_\lambda ds$$

Another view corresponding to the atmospheric situation:



$$dI_\lambda \propto \rho I_\lambda ds$$

$$dI_\lambda = -k_\lambda \rho I_\lambda ds$$

where k_λ: mass extinction cross section
Emission and multiple scattering can strengthen the intensity.

$$dI_\lambda = j_\lambda \rho ds$$

Therefore, the overall change of intensity:

$$dI_\lambda = -k_\lambda \rho I_\lambda ds + j_\lambda \rho ds$$

Assume, above term is zero (no emission or scattering contribution).

$$\frac{dI_\lambda}{k_\lambda \rho ds} = -I_\lambda$$

where k_λ: mass absorption cross section/absorption coefficient

$$I_\lambda(s_1) = I_\lambda(0)\exp\left(-\int_0^{s_1} k_\lambda \rho ds\right)$$

If the medium is homogeneous and k is independent of distance:

$$u = \int_0^{s_1} \rho \, ds$$

$$I_\lambda(s_1) = I_\lambda(0)\exp(-k_\lambda u)$$

This is Beer's Law

Beer-Bouguer-Lambert Law

The *Beer-Bouguer-Lambert Law* (also know as Beer law or Bouguer law or Lambert law) determines attenuation/extinction of radiant energy by scattering and/or absorption passing through atmosphere.

Monochromatic transmissivity T_λ

$$T_\lambda = I_\lambda(s_1) / I_\lambda(0)$$

$$T_\lambda = \exp(-k_\lambda u)$$

For a non-scattering medium, the monochromatic absorptivity A_λ:

$$A_\lambda = 1 - T_\lambda = 1 - \exp(-k_\lambda u)$$

Monochromatic reflectivity R_λ for a scattering and absorbing medium:

$$T_\lambda + A_\lambda + R_\lambda = 1$$

through a scattering and absorbing medium

Beer’s Law relates the measured intensity of solar radiation at the surface (total or monochromatic) to the one outside the atmosphere.

How can we utilize such measurements taken at the surface?

Remember the definition of the air mass concept in next two slides presented in Lecture 6 :

$$m_{act,v} = \int_0^\infty \rho \, dz. \tag{5.6.2}$$

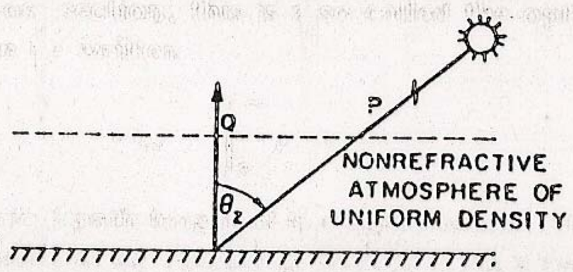
This is the mass of a substance in a vertical column of unit cross section.

The relative optical mass m_r is defined as the ratio of the optical path along the oblique trajectory to the vertical path in the zenith direction. Thus

$$m_r = \int_0^\infty \rho \, ds \bigg/ \int_0^\infty \rho \, dz. \tag{5.6.3}$$

In the foregoing, the word “air” has been deliberately avoided since attenuation of a solar beam takes place not only by dry air molecules, but also by water vapor and aerosols, etc. Therefore, Eq. (5.6.3) should be solved separately for each one of the attenuating components of the atmosphere. Optical masses for the various components are discussed below.

(a)



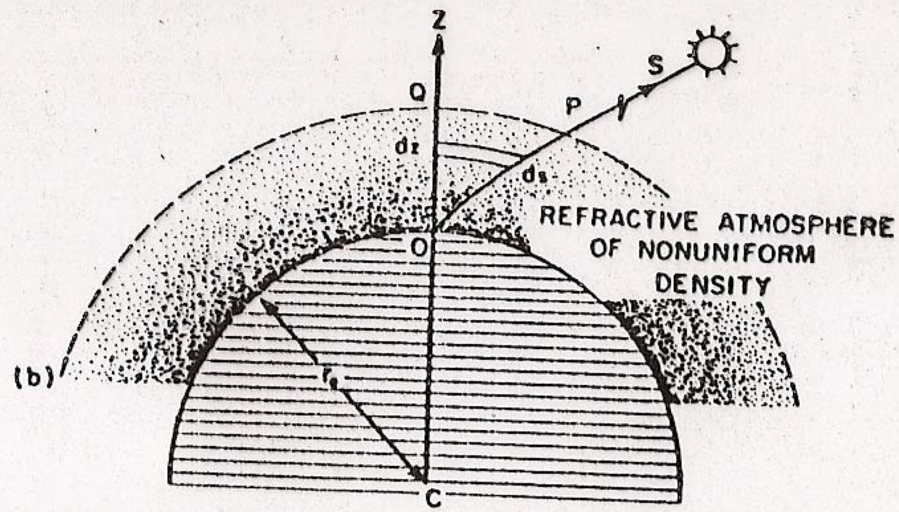


Figure 5.6.1 The trajectory of a solar ray through the earth's atmosphere. (a) Nonrefractive plane parallel atmosphere of uniform density. (b) Refractive spherical atmosphere of variable density.

Ignoring the earth's curvature and assuming that the atmosphere is nonrefractive and completely homogeneous (Fig. 5.6.1a), it can be seen that the relative optical mass applied to all the atmospheric constituents is

$$m_r' = \sec \theta_z. \tag{5.6.4}$$

We can use Beer's Law to determine extraterrestrial radiation. The procedure is labeled as "Langley Plots".

Let:

θ_0 = solar zenith angle

z_1 is the height of the station

and :

$$u = \int_{z_1}^{\infty} \rho dz$$

Accordingly, effective path length of the air mass = u sec θ_0

Beer's Law now reads:

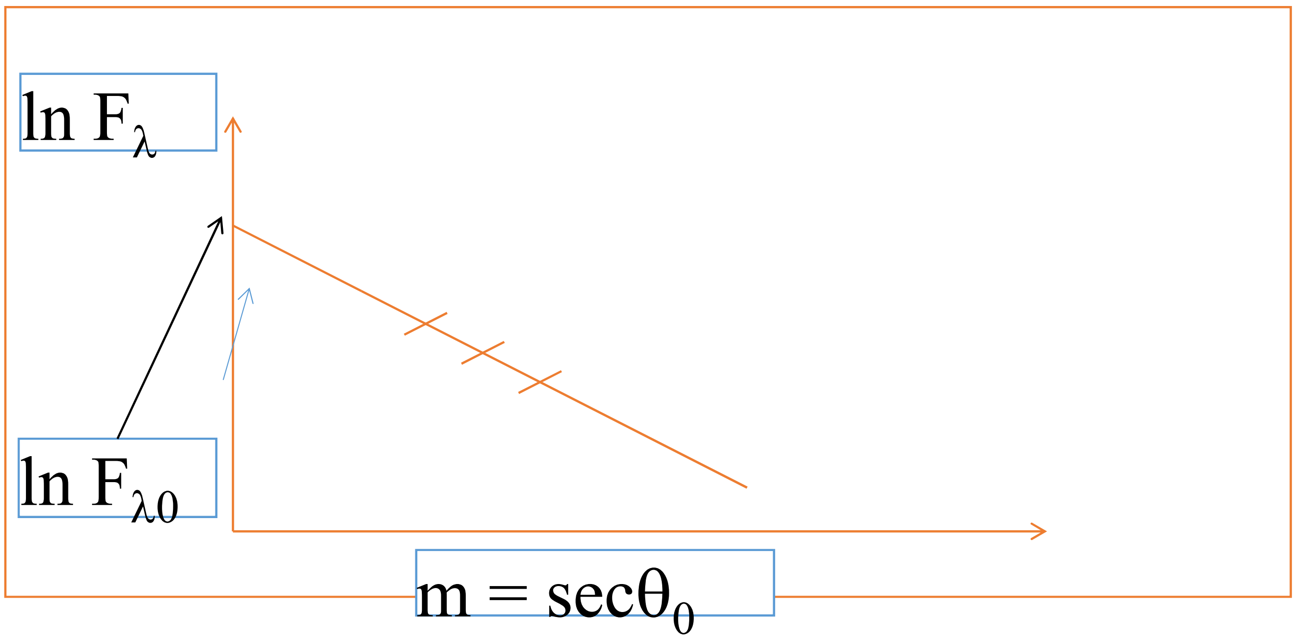
$$F_\lambda = F_{\lambda 0} \exp\left(-k_\lambda u \sec\theta_0\right) = F_{\lambda 0} T_\lambda^m$$

(For T^m see definition on slide No. 22)

$$\ln F_\lambda = \ln F_{\lambda 0} + m \ln T_\lambda$$

Langley plot

You take measurements of F_λ for different air-masses (namely, different zenith angles given by different times of day) and plot it as shown below.



Where the effective path length of the air mass = u sec θ_0. The intercept will give you $\ln F_{\lambda 0}$ which is the extraterrestrial radiation.

- If the atmospheric properties do not change during the observational period, T_λ is constant and the Langley Plot is a straight line.
- The y-interception is ln ($F_{\lambda 0}$), and the slope is ln (T_λ).
- Solar constant S:

$$S = F_0 (d/d_m)^2$$

If instrument measures F_0 (integrated over all wavelengths), possible to derive solar constant.

However, this method is not accurate to estimate S because there is energy loss:

- Absorption of UV by ozone;
- Absorption of IR by water vapor and $CO_{2;}$
- Unknown amount of diffuse radiation entering the instrument;
- Variations of k_λ and aerosols and measurement errors.

However, this method can be used for measuring monochromatic values used in sun photometers calibration *(to be explained in more detail in next lecture)*.
It is usually carried out in the early morning on a high mountain. It can be assumed that the atmosphere is stable and that the distribution of aerosols does not change much during the measurement.