

Lecture 11

IR radiation: analysis of fluxes and heating/cooling rates. Concepts of IR remote sensing.

Objectives:

1. Analysis of IR upward and downward fluxes and IR heating/cooling rates in the gaseous atmospheres.
2. Concepts of IR remote sensing.

Required reading:

L80: 7.3

Additional/Advanced reading:

Le93:22/G&Y: 6.5

1. Analysis of IR upward and downward fluxes, and IR heating/cooling rates in the gaseous atmospheres.

To calculate the IR downward and upward fluxes one needs to know:

- i) Meteorological characteristics: T, P and air density as a function of altitude (called a **vertical profile**)
- ii) The vertical profile of IR radiatively active gases, clouds and aerosols.

NOTE: In general, T, P and air density decreases with altitude (see lecture 4).

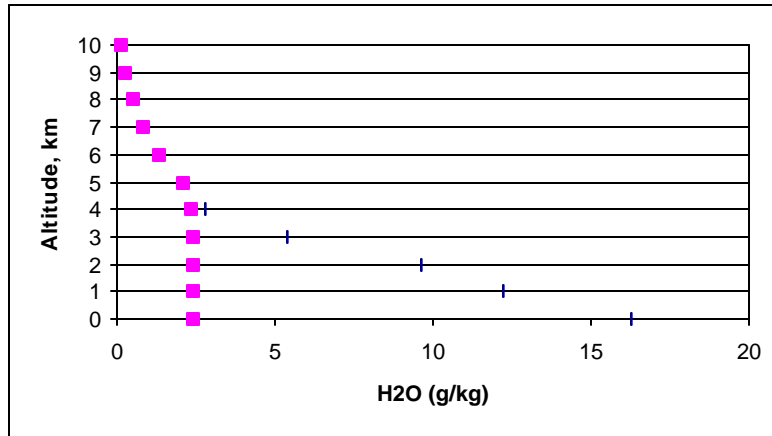
To calculate the IR heating/cooling rates one needs to know:

- i) Profiles of IR upwelling and downwelling fluxes (to calculate the profile of the IR net fluxes);
- ii) Using the profile of net fluxes and air density, one calculates the IR radiative heating/cooling rates (see Lecture 8)

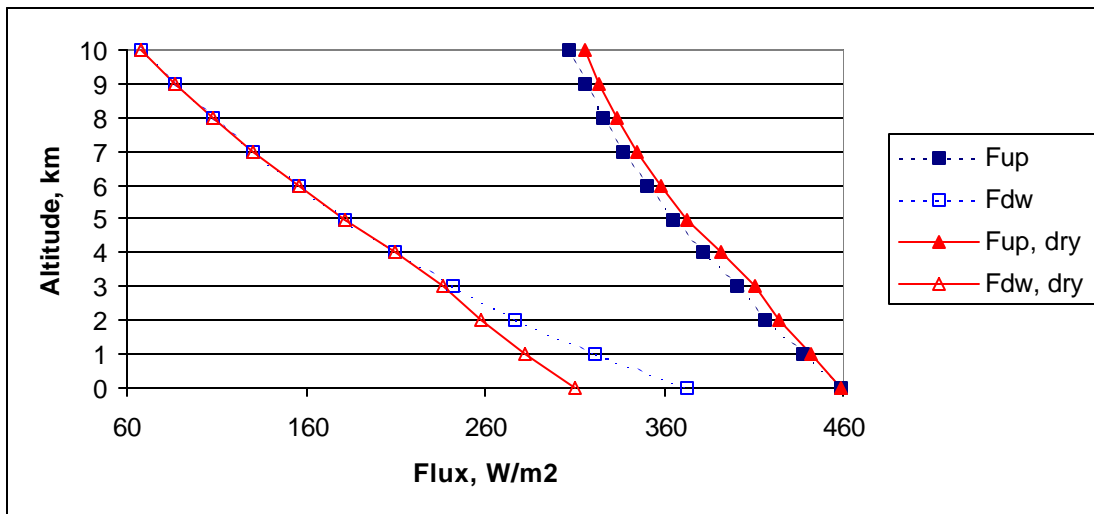
$$\left(\frac{\partial T}{\partial t} \right)_{IR} = - \frac{1}{c_p \mathbf{r}} \frac{dF(z)}{dz}$$

➤ **Effect of the varying amount of a gas on IR radiation under the same atmospheric condition**

Consider the standard tropical atmosphere and “dry” tropical atmosphere:
 same atmospheric characteristics, except the amount of H₂O

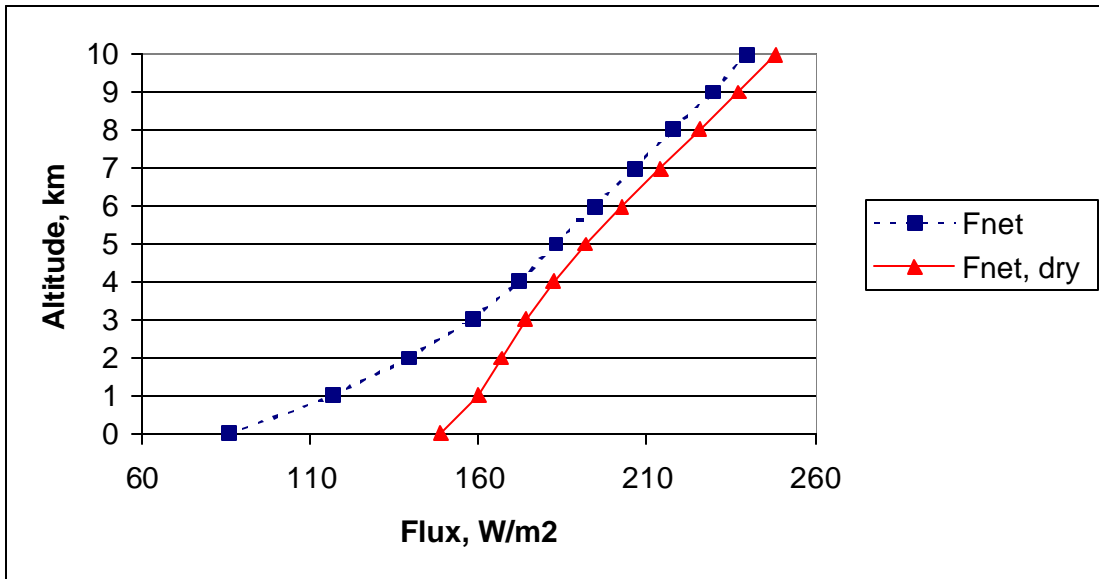


IR fluxes for tropical (dotted lines) and dry tropical atmospheres (solid lines)



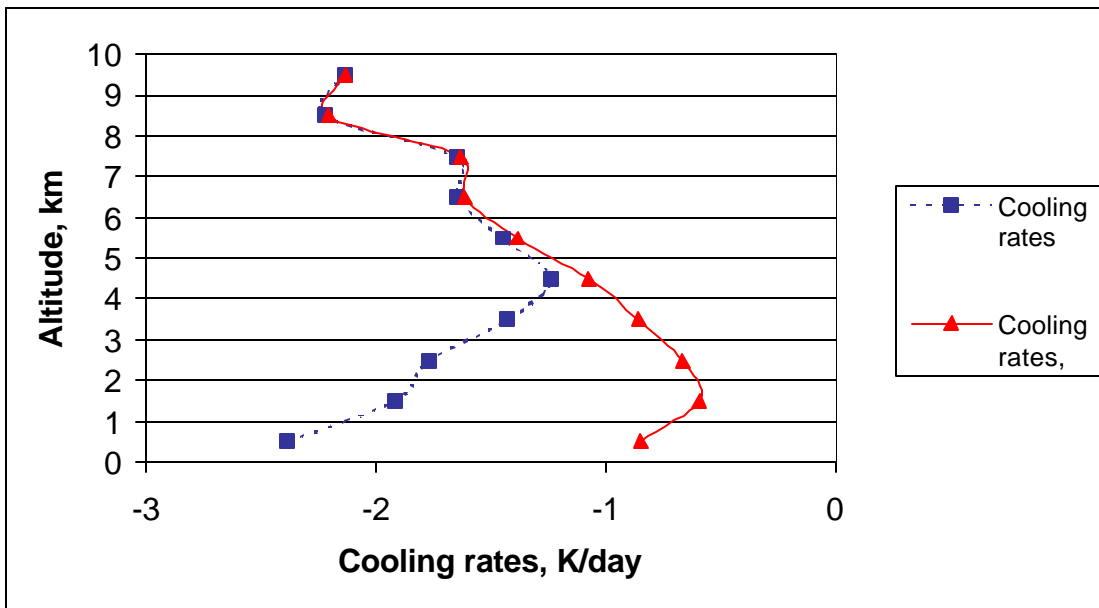
- H₂O increases in a layer ⇒ F^{\downarrow} increases because more IR radiation emitted in a layer ⇒ $F^{\downarrow}(surface)$ increases
- H₂O increases in a layer ⇒ F^{\uparrow} decreases because more IR radiation absorbed but reemitted at the lower temperature ⇒ $F^{\downarrow}(TOA)$ decreases
- Increase of an IR absorbing gas contributes to the greenhouse effect.

IR net fluxes for tropical (dotted lines) and dry tropical atmospheres (solid lines)



- The larger changes of net flux from one level to another (i.e., the larger slope of $F(Z)$ vs Z), the larger heating/cooling rates

IR cooling rates for tropical (dotted lines) and dry tropical atmospheres (solid lines)



NOTE: The largest IR cooling rates for the standard tropical atmosphere are in the surface layer.

2. Concepts of IR remote sensing.

Recall Lab1 where we have discussed the ground-based and satellite radiometric instrumentation.

Three major types of IR remote sensing:

- 1) Ground-based platforms (recall Lecture 10);
- 2) Air-borne platforms (airplanes, balloons)
- 3) Space-borne platforms (satellites)

Consider a satellite sensor measuring the monochromatic upwelling intensity (narrow viewing angle in nadir (μ is about 1)). From Eq.[6.2], we have

$$I_n^\uparrow(\mathbf{t}) = I_n^\uparrow(\mathbf{t}_1) \exp(-(\mathbf{t}_1 - \mathbf{t})) + \int_t^{\mathbf{t}_1} \exp(-(\mathbf{t}' - \mathbf{t})) B_n(T(\mathbf{t}')) d\mathbf{t}'$$

Note that $d[\exp(-(\tau' - \tau))] = -\exp(-(\tau' - \tau))d\tau'$

Then we have

$$I_n^\uparrow(\mathbf{t}) = I_n^\uparrow(\mathbf{t}_1) \exp(-(\mathbf{t}_1 - \mathbf{t})) - \int_t^{\mathbf{t}_1} B_n(T(\mathbf{t}')) d[\exp(-(\mathbf{t}' - \mathbf{t}))]$$

Recall that optical depth is defined as

$$\mathbf{t} = \int_z^\infty k_n(z') \mathbf{r}(z') dz'$$

where \mathbf{k}_n is the mass absorption coefficient and \mathbf{r} is the density.

Thus we can express the monochromatic transmission as

$$T_n(z) = \exp(-\tau) = \exp\left(-\int_z^\infty k_n(z') r(z') dz'\right)$$

Using that $\tau = 0$ at the top of the atmosphere (TOA) and $\tau = \tau_1$ at the surface $z=0$, the TOA monochromatic intensity can be written as

$$I_n^\uparrow(\infty) = I_n^\uparrow(0) T_n(0) + \int_{z=0}^{z=\infty} B_n(T(z)) \frac{\partial T_n(z)}{\partial z} dz$$

where the $dT_n(z)/dz$ is called the **weighting function**. [11.1]

NOTE: Weighting function when multiplied by the Planck function gives the upwelling monochromatic radiance contribution from a given height.

NOTE: Somewhat similar weighting functions $dT_n^f(z)/dz$ were introduced for fluxes in Lecture 8.

$I_n^\uparrow(0)$ in Eq.[11.1] represents the surface radiance $\epsilon_n B_n(T_s)$, where ϵ_n is the surface emissivity (recall Lecture 3).

Because an instrument measures an intensity in a finite spectral interval from ν_1 to ν_2 , the integration of Eq.[11.1] over this spectral interval is required. Thus we have

$$I_{\bar{n}}(0) = \frac{\int_{\nu_1}^{\nu_2} \Phi(\bar{n}, \nu) I_n(0) d\nu}{\int_{\nu_1}^{\nu_2} \Phi(\bar{n}, \nu) d\nu}$$

where Φ is the **instrumental response function** (or slit).

Therefore, Eq.[11.1] can be re-written as (assuming $\mathbf{e}_n=1$)

$$I_n^\uparrow(\infty) = \frac{1}{\int_{n_1}^{n_2} \Phi(\bar{n}, \mathbf{n}) d\mathbf{n}} \left\{ \int_{n_1}^{n_2} \Phi(\bar{n}, \mathbf{n}) B_n(T_s) T_n(0) d\mathbf{n} + \int_{n_1}^{n_2} \Phi(\bar{n}, \mathbf{n}) \int_{z=0}^{z=\infty} B_n(T(z)) \frac{\partial T_n(z)}{\partial z} dz d\mathbf{n} \right\} \quad [11.2]$$

If the spectral interval is narrow that $B_n(T)$ can be replaced $B_{\bar{n}}(T)$ and taken out of the integral over \mathbf{n} .

We can also introduce the spectral transmission as

$$T_{\bar{n}}(z) = \frac{\int_{n_1}^{n_2} \Phi(\bar{n}, \mathbf{n}) T_n(z) d\mathbf{n}}{\int_{n_1}^{n_2} \Phi(\bar{n}, \mathbf{n}) d\mathbf{n}}$$

Thus Eq.[11.2] can be re-written in a form to give a measured TOA spectral intensity in a given channel [11.3]

$$I_{\bar{n}}^\uparrow(\infty) = B_{\bar{n}}(T_s) T_{\bar{n}}(0) + \int_{z=0}^{z=\infty} B_{\bar{n}}(T(z)) \frac{\partial T_{\bar{n}}(z)}{\partial z} dz$$

Eq.[11.3] can be also expressed in the pressure coordinates: [11.3a]

$$I_{\bar{n}}^\uparrow(0) = B_{\bar{n}}(T_s) T_{\bar{n}}(p_s) + \int_{p_s}^0 B_{\bar{n}}(T(p)) \frac{\partial T_{\bar{n}}(p)}{\partial p} dp$$

- Eq.[11.3] shows that the measured spectral intensity contains information about the temperature profile and the profile of absorbing gases. Thus the multichannel measurements of TOA spectral intensities can be used to retrieve these profiles (called **inversion problem**).

NOTE: modern satellite sensors are capable of measuring the TOA intensity in about 2000 spectral bands (for instance, see Lab 1, NASA missions)

- Measurements in the 15 μm CO_2 are often used because CO_2 is well mixed in the lower atmosphere.

How to retrieve the temperature profile:

For simplicity, let's assume that $T_{\bar{n}}(p_s) = 0$ (to drop the surface contribution)

Thus

$$I_{\bar{n}}^{\uparrow}(0) \approx \int_{p_s}^0 B_{\bar{n}}(T(p)) \frac{\partial T_{\bar{n}}(p)}{\partial p} dp$$

Because $B_{\bar{n}}(T(p))$ varies from spectral channel to channel, it is common to approximate the Planck function in a linear form as

$$B_{\bar{n}}(T(p)) = c_{\bar{n}} B_{\bar{n}}(T(p)) + d_{\bar{n}}$$

where \bar{n} denotes a fixed reference spectral interval, and $c_{\bar{n}}$ and $d_{\bar{n}}$ are the empirically derived constants.

Substituting this approximation for the Planck function into the above equation for the TOA spectral radiances, we have

$$g(\bar{n}) = \int_{p_s}^0 f(p) K(\bar{n}, p) dp \quad [11.4]$$

where

$$g(\bar{n}) = \frac{I_{\bar{n}} - d_{\bar{n}}}{c_{\bar{n}}}$$

$f(p) = B_{\bar{\mathbf{n}}}(T(p))$ is **the function to be retrieved** from $g(\bar{\mathbf{n}}_i)$, $i = 1, \dots, m$, where m is the total number of the spectral radiance observations.

$K(\bar{\mathbf{n}}, p) = \frac{\partial T_{\bar{\mathbf{n}}}}{\partial p}$ is called the kernel of the Eq.[11.4]

Eq. [11.4] is a **Fredholm equation of the first kind**.

The solution of Eq.[11.4] is an ill-posed problem, since the unknown profile $f(p)$ is a continuous function of pressure but there is a finite number of observations.

NOTE: Several techniques have been proposed to retrieve the temperature profile using the multichannel measurements of TOA spectral intensities from Eq.[11.4]: direct linear inversion methods, statistical methods, numerical methods (see details in L80:7.3.3-7.3.4).