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Lecture 6

This set of Lectures will discuss
a. Solar Radiation
    1. Direct vs Diffuse Radiation
    2. Atmospheric attenuation and scattering
b. Terrestrial Radiation
    1. Long wave radiation: Stefan-Boltzmann Law
    2. Kirchoff's Law
    3. Spectral Distribution
    4. Emissivity

b. Net Radiation Balance

L6.1. Solar Radiation at the Earth's Surface

The flux density of solar radiation at the Earth's surface, on a horizontal plane, is comprised of a fraction of **direct beam** and **diffuse** radiation

$$R_g = R_{beam} + R_{diffuse}$$

The beam component is a function of the flux density of the plane normal to the sun's incoming rays and the solar elevation angle:

$$R_g = R_{normal} \sin\beta + R_{diffuse}$$

a. direct radiation

Sunlight incident at the Earth's surface passes through a gas, dust and aerosol filled atmosphere, which can absorb, reflect and scatter sunlight. Radiation reaching the Earth's surface is thereby a function of the depth of the atmosphere and the amount of attenuating particles and gases. Water vapor, ozone and aerosols are the main absorbers of solar radiation. Gas molecules and aerosols scatter light.

Measurements of sunlight, orthogonal to the solar beam rarely exceed 75% of the solar constant. **Beer's Law** is often used to evaluate the attenuation of monochromatic sunlight by the atmosphere

$$R_g(\tau,\lambda) = R_g(0,\lambda)\exp(-\tau(\lambda)\cdot m)$$

$R_g(0)$ is the radiation flux density of an aerosol free atmosphere, τ is the turbidity coefficient or optical depth and m is mass of the air, computed as the secant of the solar zenith angle or the reciprocal of the sine of the solar elevation angle ($m = 1/\sin\beta$).

$$m = \frac{P}{101.3(kPa)\cos\theta}$$

Photons encounter two types of scattering in the atmosphere **Rayleigh scattering** and **Mie scattering**. The type of scattering that occurs relates to the relation between the wavelength of the radiation and the radius of the scattering material, r:

$$s = \frac{2\pi r}{\lambda}$$

Rayleigh scattering occurs as photons strike **gas molecules** in the atmosphere, cases where the s ratio is much less than one. Rayleigh scattering pertains to **scattering** where the **medium is smaller than the lights wavelength**. This scattering is proportional to the inverse of the fourth power of the wavelength and is associated with forward and backward scattering. Shorter wavelength radiation, such as blue light is scattered more than longer radiation such as red light. If we examine scattering across the visible portion of the EM spectrum we reveal that 400 nm radiation is scattered 9 times more than 700 nm radiation $(700/400)^4$. It is Rayleigh scattering that is responsible for the blue color of earth's Atmosphere. The sun's yellow/redness color is observed because blue light is depleted from the sun's beam by the time it reaches our eyes.

Dust and **aerosols are much bigger than gas molecules** (micron) (s values between 0.1 and 50). Mie scattering pertains to these elements and is proportional to the **ratio of the molecular diameter and the inverse of the lights wavelength**, d/λ.

The optical depth is determined by integrating across the depth of the atmosphere the mass absorption (k) and scattering coefficients:

$$\tau(\lambda) = \int_0^Z \rho(k(\lambda) + s(\lambda))dz$$

In practice, the optical depth is the linear sum of the optical depths for ozone, water vapor, Rayleigh scattering and aerosol Mie scattering:

$$\tau(\lambda) = \tau_{o_3}(\lambda) + \tau_{h_2o}(\lambda) + \tau_{Rayleigh}(\lambda) + \tau_{aerosols}(\lambda)$$

Long term and high quality measurements of atmospheric transmission are becoming more important, to assess another aspect of global change.

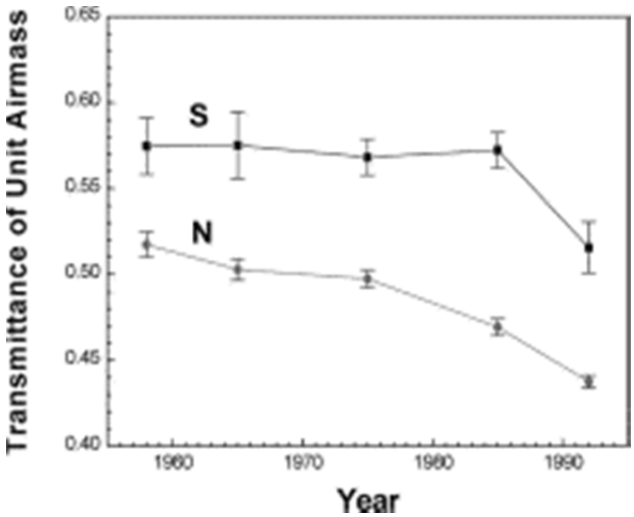


Figure 1 Evidence of Solar Dimming (Stanhill and Cohen, 2001)

b. diffuse radiation

Diffuse radiation is radiation from the sky, exclusive of the sun. There is a component of diffuse radiation on clear and cloudy days.

The amount of diffuse radiation reaching the surface will depend on cloud type, the sun's elevation angle, and the amount of reflected surface radiation that is reflected downward again. The fraction of diffuse light is increasing as more aerosols are loaded in the atmosphere due to pollution or after volcanic eruptions. Stanhill and Cohen (2001) have reported that the transmissivity of the atmosphere is decreasing. They report that the maximum rate of decrease in solar radiation was -1.7 W m^{-2} per year, or -1.2% per year. The location of this trend was centered around 35°N, a region with high insolation, fossil fuel use and population.

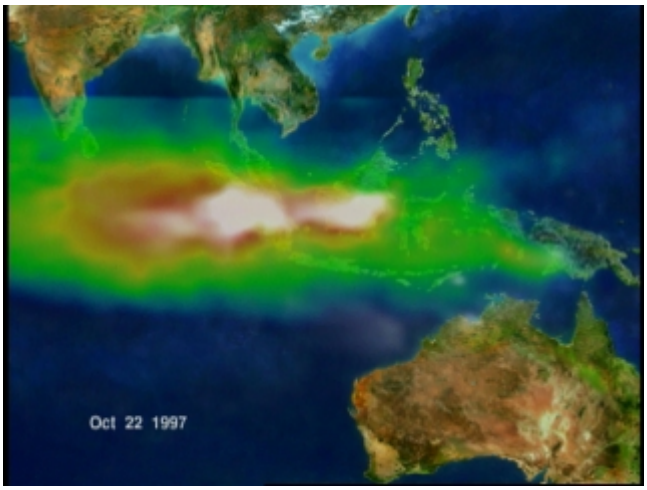


Figure 2 http://www.gsfc.nasa.gov/gsfc/earth/environ/indo/cu1022.jpg

The spectral composition of diffuse light is altered from that in the sun's direct beam. The mean spectrum of diffuse light in a clear sky is shifted towards shorter wavelengths and has a mean of 450 nm. Under clouds, scattering cause the sky to be grayish.

The zonal distribution of sunlight under a clear sky can be described with the **uniform overcast sky distribution (Wallace and Hobbs, 1977)**:

The amount of energy available to the hemisphere can be evaluated using solid geometry. First we start with the definition of the solid angle, ω, in terms of the zenith and azimuth angles:

$$d\varpi = \sin\phi d\phi d\theta$$

and its integral over the upper hemisphere

$$\int_0^{\pi/2} \sin\phi d\phi d\theta \int_0^{2\pi} d\theta = 2\pi$$

The flux density of diffuse radiation contributed across the sky is

$$E = \int_0^{2\pi}\int_0^{\pi/2} L\cos\phi\sin\phi d\phi d\theta$$

$$E = 2L\int_0^{\pi/2} L\cos\phi\sin\phi d\phi$$

So for a given sky sector the fractional radiation is:

$$\frac{R_d(\gamma)}{R_d} = 2\cos\gamma\sin\gamma = \Gamma(\gamma)$$

Under **cloudy skies** the standard overcast sky distribution is used.

$$\frac{R_d(\gamma)}{R_d} = \frac{6}{5}(1+\sin\gamma)\cos\gamma\sin\gamma$$

It shows that the sky near the zenith is 3 times brighter than near the horizon.

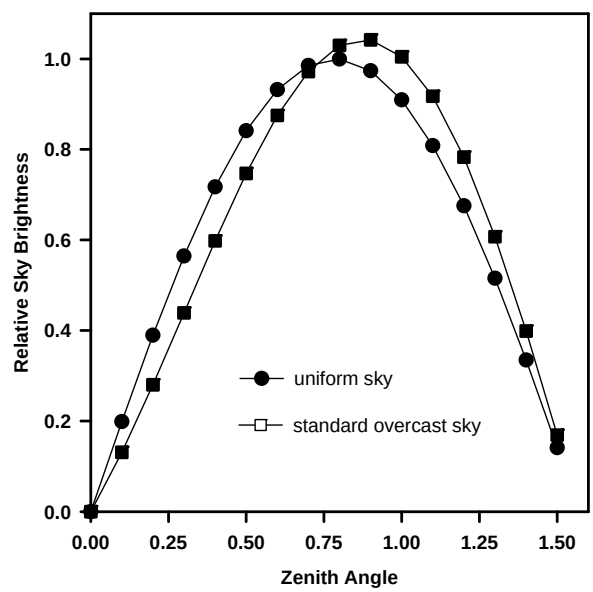


Figure 3 Calculations of sky brightness for a uniform sky and standard overcast sky

L6.2. Long wave Emittance

Black body radiation is another important concept to grasp. Photons emitted by a energy transition emit photons at a single wavelength. A similar relation holds for photons and energy that are absorbed. When spectral lines merge, as with an infinite combination of absorption and emission lines, the spectra becomes continuous. A black body radiator thereby is defined as a surface that perfect absorbs and emits radiation. In reality, no surface truly exists. But many surfaces are close to being black body radiators for infrared radiation.

We can define **emissivity**, the fraction of black body emittance at a given wavelength by a material. If the surface is a true ***black body*** then **Kirkoff's Law** applies:

$$\alpha(\lambda) = \varepsilon(\lambda)$$
$$\rho(\lambda) = 0$$
$$\tau(\lambda) = 0$$

For a black body surface, absorptivity equals emissivity in the long wave band and they equal one. Reflectance and transmissivity are zero.

Not all surfaces are perfect black body emitters (ε=1).

$$\alpha(\lambda) = \varepsilon(\lambda)$$
$$\rho(\lambda) = 1 - \varepsilon(\lambda)$$
$$\tau(\lambda) = 0$$

Substances such as aluminum have low emittance factors. In this situation the Stefan-Boltzmann law is redefined as:

$$L\uparrow = \varepsilon\sigma T^4$$

Most plants and soils of interest have high emissivities (> 0.9). A list of values is presented in the following Table.

Table 1 List of Infrared Emissivities (Campbell and Norman, 1998)

Surface	Emissivity
Plant leaves	0.94-0.99
Glass	0.90-0.95
Aluminum	0.06
Aluminum paint	0.30
Soil	0.93-0.96
Water	0.96

For non-black bodies we can still assume transmission is zero. To calculate reflectivity we use:

$$\rho(\lambda) = 1 - \varepsilon(\lambda) = 1 - \alpha(\lambda)$$

So the infrared reflectivity of water, that has a 0.96 emissivity, is 0.04.

L6.4 Terrestrial Radiation

According to **Kirchoff's law**, gases that absorb radiation emit radiation. Gases in the atmosphere that absorb and emit radiation are water vapor, ozone and carbon dioxide. Nitrogen and Argon absorb little if no radiation. Oxygen, though it is considered to absorb no radiation, does disassociate by absorbing ultraviolet radiation to form ozone.

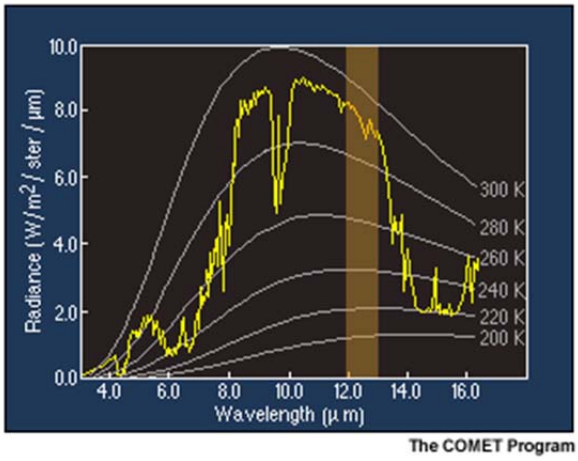


Figure 4 meted.ucar.edu/satmet/goeschan/ media/graphics/sm5cm92.gif

Several empirical functions have been developed to compute downward longwave radiation for cloudless skies (Monteith and Unsworth, 1990).

$$L\downarrow = \varepsilon_a \sigma T_{a,K}^4$$

where the longwave flux density and the emissivity are a function of air temperature and humidity, respectively.

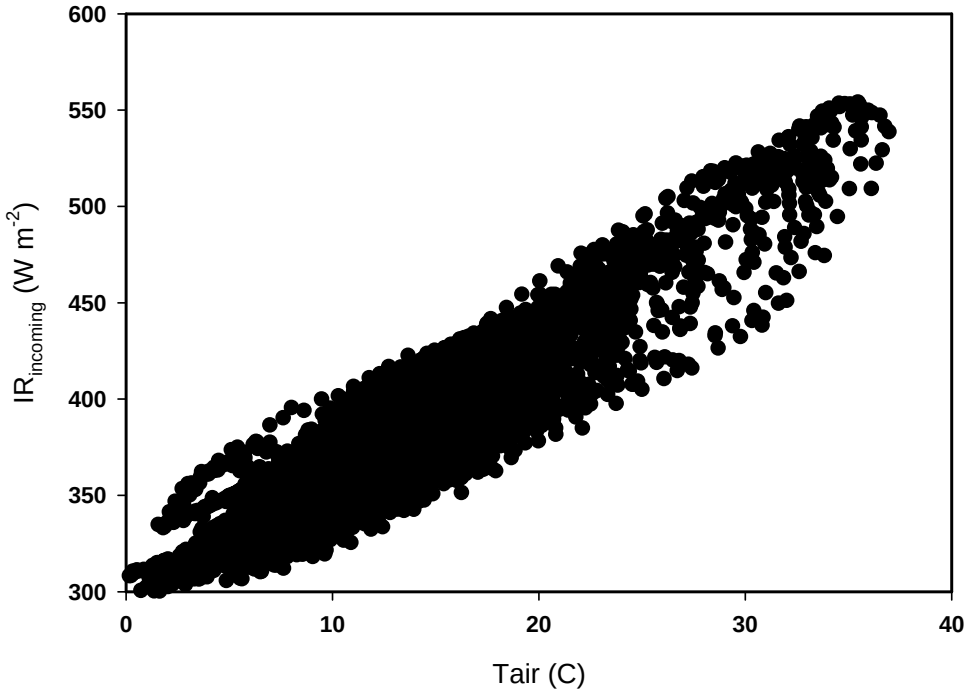


Figure 5 Measurements of incoming longwave radiation, near Ione, CA.

Based on work by Monteith and Unsworth, the emissivity can be estimated as a function of precipitable water in the atmospheric column, p.

$$\varepsilon_a = a + b(\ln(p) + 0.5)$$

Brutsaert recommends an algorithm for clear sky emissivity that is a function of the vapor pressure of the atmosphere

$$\varepsilon_{sky} = 1.72\left(\frac{e_a}{T_a}\right)^{1/7}$$

where ea is vapor pressure (kPa) and Ta is air temperature (K).

On the basis of empirical field data, numerous algorithms exist for estimating downwelling longwave energy flux density. Monteith and Unsworth advocate:

$$L\downarrow = -119 + 1.06\sigma T_a^4$$

where T is in degrees Kelvin

With further manipulation we can arrive at the equation:

$$L\downarrow = 213 + 5.5T_a$$

where T is in Celsius.

Evaluation of these functions show that longwave radiation ranges between about 150 and 450 W m^{-2} in the temperature range of -10 and 40 C.

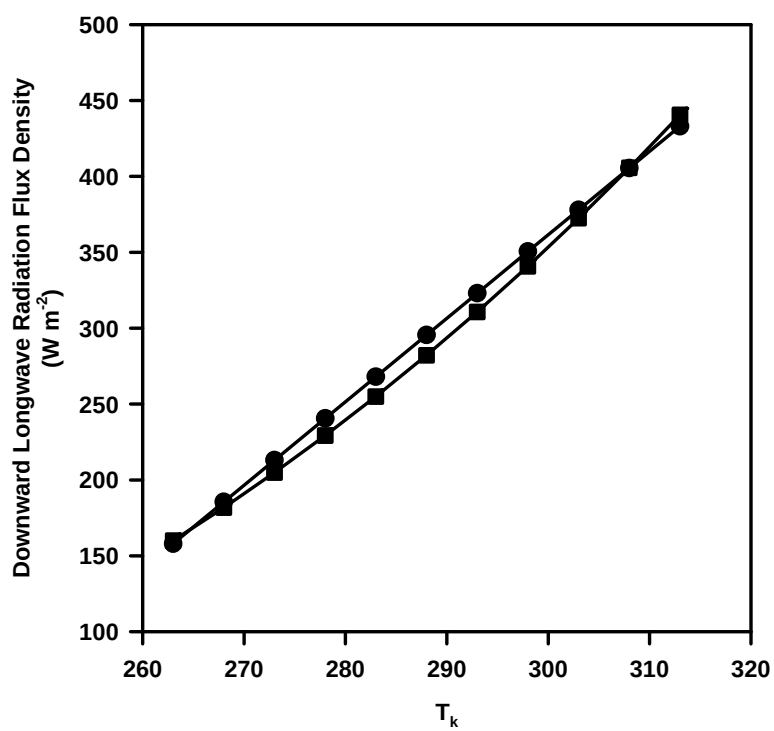


Figure 6 Model computations of Infrared flux density as a function of temperature

Other functions are attributed to Swinbank (Swinbank, 1963)

$$L\downarrow = 5.31 \cdot 10^{13} T_a^6$$

The upward emitted long wave radiation is a function of the surface temperature (to the fourth power) and the downward longwave radiation that is reflected upward.

$$L\uparrow = \varepsilon\sigma T_s^4 + (1-\varepsilon)L\downarrow$$

L6.3 Net Radiation Balance

The surface energy budget provides the conceptual guide for investigating energy exchange over a landscape. The net radiation (R_n) absorbed by a forest, crop, fen or lake is equal to the sum of incoming short wave (R_g) and long wave radiation ($L\downarrow$) minus the reflected shortwave and emitted long wave radiation.

$$R_n = (1-\overline{\alpha})R_g + L\downarrow - L\uparrow$$

$$R_n = (1-\overline{\alpha})R_g + \varepsilon L\downarrow - \varepsilon\sigma T_s^4$$

Before we close, it will be instructive to use some of the principles of the radiation, at hand, to examine the radiation balance of earth. As biometeorological studies can contribute to our understanding of Earth's greenhouse effect.

Case 1. Case with no atmosphere.

Equilibrium balance between radiation intercepted by Earth and that radiated back into space.

$$\pi \cdot r_{earth}^2 (1-\alpha) S^* = 4\pi \cdot r_{earth}^2 \sigma T_k^4$$

$$\frac{(1-\alpha)S^*}{4} = \sigma T_k^4$$

S*= 1366 W m^{-2}
α=0.30, mean Earth Albedo

$E=241\ W\ m^{-2}$
$T_{rad}=255K$, mean radiative temperature of the Earth

2. Radiative balance over planet with atmosphere, the Zero Dimensional Case

Everyday experience, the presence of liquid water and a network of climatological temperature sensor tells us that the surface temperature is way above 255 K (-36 C). Greenhouse gases partially close the atmospheric window and trap heat more efficiently at the surface and in the lower atmosphere. To balance this effect, the surface and lower atmosphere must warm and emit thermal radiation at a greater intensity.

'Greenhouse' Effect is caused when the atmosphere absorbs thermal radiation in the spectral regions where the atmosphere is opaque. In accordance with the conservation of energy, the amount of energy absorbed equals that emitted. With the case of an atmosphere overlying the earth, approximately half of the emitted radiation is re-directed towards the surface and the other half is lost to space.

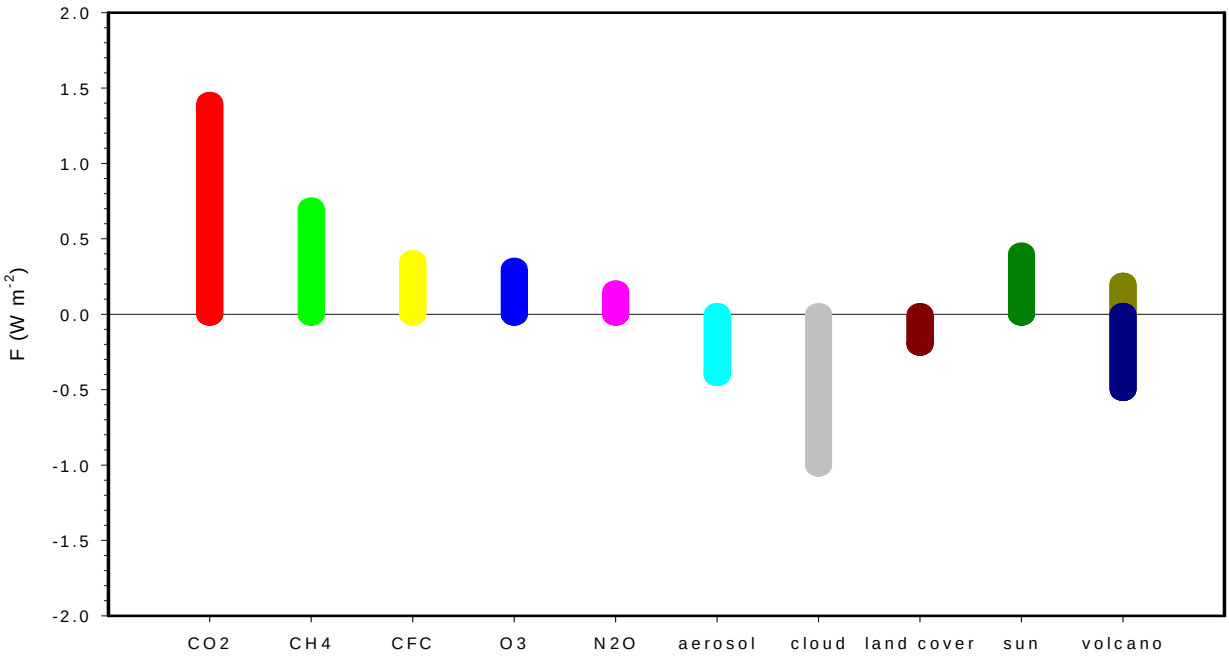
Top of atmosphere, 342 W m^{-2} received from sun, 105 W m^{-2} is reflected for a net of 237 W m^{-2}. Yet 390 W m^{-2} of radiation is received and emitted at Earth's surface since its surface temperature is 288K, causing long wave radiation to peaks at about 10 um. To enable this much energy to be received at the surface the atmosphere must absorb and re-radiate 153 W m^{-2}, the difference between the surface emission and that lost to space

The direct radiative forcing of long-lived trace gases is about 2.45 W m^{-2}

CO_2 forcing: 1.56 Wm^{-2}
CH_4 forcing: 0.47 Wm^{-2}
N_2O forcing: 0.14 W m^{-2}

The direct radiative forcing of CFC and HFC is about 0.25 W m^{-2}, but the net forcing is 0.1 W m^{-2} due to a negative feedback associated with ozone depletion in the stratosphere.

Tropospheric aerosols, due to fossil fuel and biomass combustion, have a negative forcing of about 0.5 W m^{-2}. These aerosols are short-lived and easily washed out of the atmosphere. This forcing adjusts rapidly to changes in emissions.



adapted from J. Hansen

DISCUSSION QUESTIONS/POINTS TO PONDER

Is it a coincidence that our vision and the waveband for photosynthesis evolved to utilize light in the peak spectral range of the sun?

SUMMARY POINTS

- Longwave emission of energy is proportional to its emissivity, as defined by Kirkoff's Law
- The net radiation balance is a equal to the sum of incident solar radiation minus the fraction reflected (albedo times incoming solar radiation) plus incoming longwave energy minus the energy emitted from the surface that is proportional to its temperature to the fourth power.

Endnote References

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