THE EXOSPHERIC HEAT BUDGET

What determines the temperature **on** earth ? In this course we are interested in quantitative aspects of the fundamental processes that drive the earth machine. We will mainly look at global <u>mean values</u> of heat budgets and not too much at the zonal variation. Important variables are 1. solar energy input and spectral distribution; 2. atmospheric albedo; 3. surface albedo; 4. concentrations of greenhouse gases; 5. concentrations of short wavelength absorbers

The solar radiation energy at one astronomical unit (distance of earth to sun) is called the insolation constant S, about 1367 W/m^2; we can assume that all that radiation arrives at the earth as a parallel beam (earth is small within the solar system context). The earth intercepts as a disk with radius R on the order of 10^{17} watts, but that amount of energy is "smeared out" over its spherical surface. We can recalculate the <u>average</u> terrestrial insolation constant S* as
$\mathbf{S^* = S \times \pi R^2 / 4\pi R^2}$, where R is the earth radius. The amount of energy intercepted equals S times the surface of a disk with radius R; this energy is distributed over the surface of the earth sphere. The value of S* is S/4 = 341 W/m^2, obviously averaged over a 24 hour period. Physicists know a lot about radiative processes so a few physics laws follow:

1. Planck's Law states that for emission from a blackbody at temperature T the amount of energy per wavelength band can be calculated according to a complex function with λ and **T** (T is the absolute temperature). The full expression is

$$\mathbf{B_\lambda(\lambda, T)\, d\lambda = \{(2hc^2) / \{\lambda^5\, [e^{ch/k\lambda T} - 1]\}\, d\lambda}$$

2. The Stefan-Boltzman law states that the emitted radiative energy per 'surface unit' scales with $\mathbf{T^4}$ (in degrees K). This is the integrated form over all λ of Planck's Law. $\mathbf{E = \sigma T^4}$, where σ is the Stefan Boltzman constant.
3. Wien's Law gives the wavelength with the strongest intensity for a given blackbody temperature: $\mathbf{\lambda_{max} . T = 3\ 10^6}$ (λ in nm) This law is derived by setting "the first derivative of Planck's Law with respect to λ" equal to 0.
4. Kirchhof's law says that for a substance the "tendency to emit is equal to the tendency to absorb" and I add that it both happens with the same λ.
5. The Beer-Bouguer-Lambert Law describes absorption for radiative processes: $\mathbf{(I_\lambda{}^s) = (I_\lambda{}^o)\, e^{-k\,\rho\, s1}}$, where for a given wavelength λ the absorbed intensity ($\mathbf{I_\lambda{}^s}$, at pathlength s) is equal to the original intensity ($\mathbf{I_\lambda{}^o}$) multiplied by the exponential term, where k is the absorption coefficient, ρ the gas density (or another parameter for concentration of the absorber) and s1 the pathlength.

$\mathbf{(I_\lambda{}^s) / (I_\lambda{}^o)}$ is also known as the transmissivity, and equals $\mathbf{e^{-\mathrm{INT}[k\lambda\ \rho\ ds]}}$

where the term under the integral is also known as the optical depth. The integration has to be done if the gas concentration varies over the pathlength s1. The same relation can be written for scattering instead of absorption, and k is then the scattering coefficient. A combined k* value is in common use, which takes care of both the scattering and absorption in a medium.

The on-line routines show the observed solar spectrum, and from Boltzman's Law we learn (after application of the $1/R^2$ relationship) that the observed S value is produced by a solar temperature of about 5600 K, presumably the temperature at the outside surface of the sun. From Wien's law we then get that the maximum intensity wavelength is about 550 nm; Planck's law will give us the main wavelength range: about 100 – 3000 nm, or from UV into IR.

Important processes in the radiation balance of the earth are scattering, absorption, and reflection of the solar radiation in the atmosphere, and reflection + absorption by the solid earth. The observed radiative flux from the earth has a spectrum largely in the IR : the most intense wavelength is at about 1100 nm (range: 400-4000 nm), and the radiative terrestrial temperature is about 255 K (-18 C!), a temperature found at about 5 km altitude. This radiative temperature can be derived from the steady state condition (incoming radiation = outgoing radiation) $\mathbf{(1-\alpha)S = \tau\ T^4}$ where α is the total planetary albedo and τ the Stefan-Boltzman constant. The earth surface temperature T_o is 288 (15 °C) and we can rephrase the above expression $\mathbf{(1-\alpha)S = T_e\tau\ T_o^4}$ where T_e is the effective transmissivity, which has an average value of about 0.6 for the earth (but varies strongly with wavelength). We are thus not seeing much of the solid earth long-wavelength radiation at the outside of the atmosphere, but instead we see radiation that comes from the upper troposphere. This is a direct result of the actions of the greenhouse gases which absorb the long-wavelength from the solid earth radiation, and then re-radiate to outer space at their own temperature! The modern greenhouse effect raises the earth surface temperatures by 18+15 =33 C. **But see my discussion elsewhere on this simple statement!**

The radiation budget of the atmosphere+ solid earth is probably known to all of you. The total heat balance at the earth surface is not purely radiative because, simply said, $Q_{abs} - Q_{rad} = +$. Thus more solar heat is absorbed at the earth surface then is given off by radiation. To maintain thermal balance, the difference is transported by convection. The table below gives the proportions of these fluxes in W/m^2. Remember S* = 239 W/m^2.

	RADIATION			**CONVECTION**	
	short λ	**long λ**	**net**	**sensible**	**latent**
atm	**65**	**-176**	**-111**	**19**	**92**
earth surf	**174**	**-63**	**+111**	**-19**	**-92**
total	**239**	**-239**	**0**	**0**	**0**

About 60% of the absorbed short-wavelength radiation (solar) at the surface of the earth is transported through advection with water vapor and as convection with sensible heat, the rest is radiated away.

All these considerations are for annual, whole-earth radiative balances. Now we are ready to look at zonal differences. The surface as well as the atmospheric albedo vary strongly as a function of latitude: snow reflects a lot more light than corn fields! In addition, the angle of incidence of the solar radiation varies as a simple function of latitude: the incoming energy is smeared over a larger surface area at higher latitudes and the albedo effects is also sensitive to angle of incidence. These are two reasons that the polar regions will have a much smaller absorbance of solar radiation. The terrestrial radiation also varies as a function of latitude, largely a reflection of the temperature. We can now subtract the terrestrial radiation from the net incoming radiation *per* latitude band and the result is a regional account of the radiative balance. We can approach this from the empirical side based on satellite measurements. The poles are losing heat, with the N pole losing more heat than the south pole. The oceans gain more heat than the landmasses, which sets up a condition for heat transfer from the oceans to land, especially during the winter. The total amount of heat transport from low latitudes to higher latitudes can be as high as 6×10^{15} W. This heat transport occurs through atmospheric currents that carry sensible heat and latent heat with water vapor, and ocean currents that carry largely sensible heat and ice as negative latent heat. Much of climatology deals with the distribution of heat over earth, which is obviously strongly tied to this zonal radiative balance. These balances can be upset by volcanic eruptions, cloud cover (e.g., modified from "seeds" from pollution) and of course, concentrations of greenhouse gases.

The Greenhouse Effect

Let us now look more in detail at the nature of the greenhouse effect. The incoming solar radiation has only a small amount of energy in the infrared bands, but the terrestrial radiation is largely within the infrared wavelengths. We can thus argue that a large part of the incoming solar radiation is transmitted through the atmosphere (apart from scattering, reflection, and short-wavelength absorption, etc) whereas a substantial part of the outgoing long-wavelength terrestrial radiation is absorbed. Absorption of the short (ultraviolet) wavelengths leads to electronic transitions, whereas infrared absorption leads to changes in the vibrational and rotational spectra of the molecules. Dimeric molecules are radiatively neutral but "things" with several arms can either change their rotational or vibrational energies. Some of those states are indicated in fig. 5, and each vibrational or rotational absorption occurs at a given wavelength. The likelyhood of absorption of a photon with the appropriate wavelength for a substance is labeled the <u>absorption cross section</u>.

Many molecules have several absorption bands, related to different vibrations and rotations. Important terrestrial greenhouse gases are given in table

2. The absorption spectrum of the atmosphere is shown in figure 2. We can express this also in T_e values for each wavelength (plotted here as the inverse of T_e, % absorption in fig. 2). When we look at T_e on a *per* wavelength basis we can see that some windows are very "dirty", with a T_e close to zero. These dirty windows are caused by the infrared absorption by CO_2 and H_2O. There are "open" windows in the spectrum and that is where the bulk of the terrestrial radiation escapes through to outer space (the 6 % from figure 3). It is intuitively clear that absorbers that act within these relatively clean windows will have a proportionally large effect on the total atmospheric transmissivity. Important absorbers in these open windows are methane (CH_4) and nitrous oxide (N_2O), which have overlapping absorption bands.

We are now in the position to investigate the radiative effect of changes in the concentrations of greenhouse gases, the so-called radiative forcing of the greenhouse gases. The radiative forcing factors for some common greenhouse gases are given in table 2. The CO_2 and H_2O windows are already dirty, and so a little bit more CO_2 will not dramatically influence the radiative balance. The CO_2 radiative forcing thus scales with the logarithm of the concentration. The CFC's have several absorption bands and absorb in a virtually clean window of the natural atmosphere: their greenhouse forcing is therefore on a mole by mole basis very large compared to CO_2. Sometimes the greenhouse forcing is thus listed as the CO_2 equivalent greenhouse effect of a gas as applied to the modern atmosphere (Table 3). It is a common error in many low-level textbooks and newspaper articles to argue that methane and the CFC's have a much larger influence on the radiation balance than CO_2 only because of a compound-intrinsic property (e.g., because they are much better absorbers that supposedly have a larger cross section and/or more absorption bands). The fact is, they absorb in a region of the spectrum where no other natural compounds absorb, and especially for the synthetic CFCs, the influence of each newly released molecule thus counts heavily. The CFCs do indeed have absorption bands with intensities about 10 times greater than CO_2 absorption at 15 μm (the main CO_2 absorption band), but compare this with the CFC factors in table 3. This overview shows that the calculation of the total greenhouse effect with a changing atmospheric composition is not an easy thing, partially because of band overlap, and non-linear effects also play a role.

A crude and semiquantitative way of assessing the temperature effect of changes in either greenhouse forcing, albedo or solar intensity is as follows: take the log of the equation on the radiative steady state condition given above and then differentiate all terms (and we use f for T_e), leading to

$$\delta f/f + 4.\ \delta T/T = \delta S/S - (\delta\alpha/1-\alpha)$$

It can be derived that when all else remains the same, a relative decrease in the atmospheric greenhouse effect of about 1 % ($\delta f/f$ = 0.01) gives a temperature change of about 0.7 K, and the same is true for a 1 % increase in solar intensity (S). An increase in albedo of 1 % would lead to about 1 K cooling.

These are approximate figures, neglecting dependencies of albedo on greenhouse gas concentrations (e.g., water vapor and clouds). Variations in S can be partially related to sunspot activity, which has a cycle of about 11 years, but longer periods with low sunspot activity are also known (e.g., 17th century). The earth orbit is not exactly circular so at perihelium (July) the S value is about 16 W/m^2 larger than at apehelium (December), a difference of about 7 %.

It is obvious from the foregoing that the greenhouse effect retains some of the radiated energy from the earth surface in the atmosphere: unidirectional radiation from the earth surface (outwards) is absorbed and re-radiated in all directions, including back to the surface. This re-radiated energy can be absorbed again and re-radiated once more. The earth surface will thus warm up as a result of the greenhouse effect. This radiative transport back and forth will not change the temperature of the atmosphere, however, which is a function of the kinetic energy of the main gas molecules, N_2 and O_2, which are neutral in radiative sense (do not absorb or emit long-wavelength fotons). How is the energy transferred from the trace greenhouse components to the main mass of gases? This is another one of those issues that is rarely addressed in detail in most textbooks. Realize that absorption of the quants of energy by greenhouse gas molecules only raises their <u>internal energy</u>. That process is not changing either their temperature or that of other gas molecules. The transfer of energy must partly occur in the short timespan between absorption and re-emittance of the infrared quants through molecular collisions. In the dense troposphere collisions are manifold and so a large amount of energy can be transferred to the "inert" main gas molecules. In addition, there is the convective heat transport from the earth surface into the atmosphere, although the actual heat exchange between earth surface and atmosphere is conductive, a relatively slow process.

The absorption spectra of short-wavelength absorbers, largely O_2, N_2, O_3, and strange N species are shown in figure 6, with the atmospheric penetration depths of the very short wavelengths at the bottom.