Today's Lecture (Lecture 2): Radiation

Reference

Hartmann, Ch. 3

Peixoto and Oort, Ch. 6 (much more detailed than our treatment)

Peixoto and Oort, Sec. 3.1, 3.2, 3.4, 3.5 (in preparation for next week); skip discussion of oceans until one week later

2 – The climate system

1. Introduction

2. The climate system
   - 2.1 Radiation
   - 2.2 Atmosphere: fundamental equations
   - 2.3 General circulation of the atmosphere
   - 2.4 Ocean
   - 2.5 General circulation of the oceans
   - 2.6 Land, biosphere
   - 2.7 Cryosphere

3. Internal variability

4. Forcing and feedbacks

5. Anthropogenic climate change

Reference

Hartmann, Ch. 3
Peixoto and Oort, Ch. 6 (much more detailed than our treatment)
Peixoto and Oort, Sec. 3.1, 3.2, 3.4, 3.5 (in preparation for next week); skip discussion of oceans until one week later

The climate system

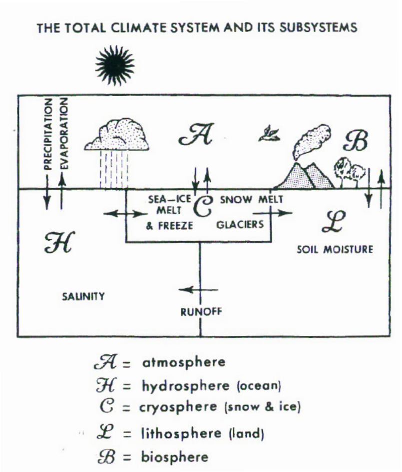


Figure: Peixoto and Oort

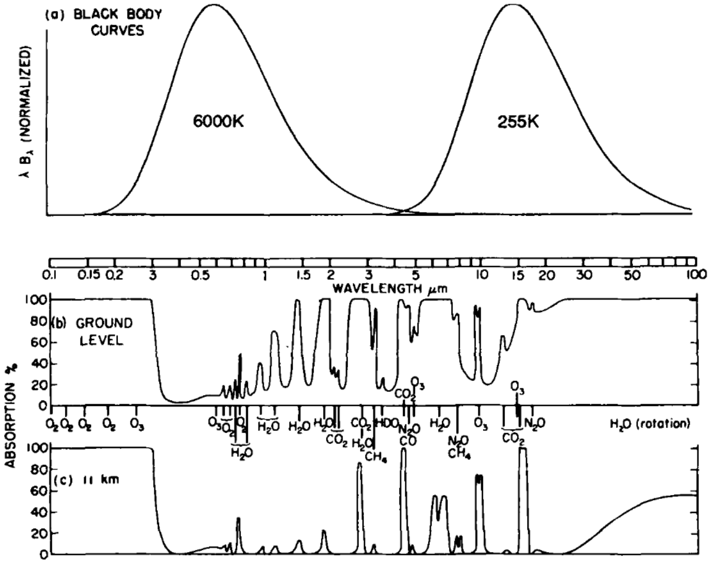
Boundary of the climate system

We consider the "top of the atmosphere" (TOA) the boundary of the climate system. The energy balance at the TOA determines whether the climate system is in equilibrium.

2.1 – Radiation

- Absorption and emission by surface and atmosphere
- Greenhouse effect
- Responses of the climate system to radiation: convective adjustment, meridional heat transport
- Summary of radiative fluxes

Emission and transmission by Sun, Earth and atmosphere



- Terrestrial (“longwave”, *R*) and solar (“shortwave”, *S*) spectra have nearly no overlap
- Ultraviolet absorption: O_3 (main stratospheric absorber)
- Visible absorption (very little), Near IR (H_2O, CO_2)
- Terrestrial: H_2O (main tropospheric absorber), CO_2, CH_4, O_3, N_2O
- Only “window” for terrestrial radiation is between 6.3 and 12 μm
- CO_2 has a strong absorption peak at 15 μm, near the maximum of terrestrial emission, where absorption is saturated
- Atmosphere is ≈ transparent to solar, ≈ opaque to terrestrial radiation

Figure: Goody and Yung (1989)

Thermal emission

We need to distinguish between two forms of matter with very different radiative properties:

Solids and liquids

Solids and liquids are approximately *black bodies*. A black body is one that absorbs all incident radiation and re-radiates according to its temperature with radiative flux (energy per unit time per unit area)

$$R = \sigma T^4 \tag{2.1}$$

The equivalence is approximate because solids and liquids do not absorb all incident radiation. However, their departure from perfect absorption is a slowly varying function of wavelength. The constant of proportionality in the thermal part of the spectrum is called *emissivity*, and the radiative flux is then

$$R = \epsilon\sigma T^4 \tag{2.2}$$

(with $0 \leq \epsilon \leq 1$). The emissivity of most non-metallic surfaces is between 0.9 and 1.

Gases

Gases are terrible black bodies. Absorption and emission occurs in narrow wavelength bands corresponding to energy level transitions. Therefore, any realistic flux calculation requires (the computation equivalent of) an integral over wavelengths. (We will nevertheless perform some crude calculations below that treat the atmosphere as a black body — but that have illustrative results.)

Planck's law

As a result of photon quantum statistics, the spectral radiance of a perfectly absorbing body (i.e., the energy emitted per unit time per unit area per unit solid angle per unit wavelength) is a function only of temperature (*Planck's law*):

$$B_\lambda(T)\,d\lambda = \frac{2hc^2}{\lambda^5}\frac{1}{\exp\left(hc/\lambda k_B T\right) - 1}\,d\lambda \tag{2.3}$$

(note: λ is wavelength here, not longitude). The total radiance (i.e., the energy emitted per unit time per unit area per unit solid angle)

$$B(T) = \int_0^\infty B_\lambda(T)\,d\lambda \sim T^4 \tag{2.4}$$

We often idealize the planetary surface, a layer of the atmosphere, a layer of cloud, etc., as an infinite plane. In this case, the *hemispheric* radiative flux (i.e., the energy emitted per unit time per unit area *upward or downward*) is given by

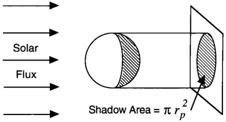
$$R = \int_0^{2\pi} d\phi \int_0^{\pi/2} B(T)\cos\theta\sin\theta\,d\theta = \pi B(T) = \sigma T^4 \tag{2.5}$$

(where $d\phi\,d\cos\theta$ in this context is a solid-angle element – nothing to do with latitude or potential temperature).

$$\sigma = \frac{2\pi^5 k_B^4}{15h^3c^2} = 5.670 \cdot 10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}} \tag{2.6}$$

is the Stefan–Boltzmann constant.

Greenhouse effect



The global average solar flux, averaged over day and night and over latitudes, can be calculated by a cute trick:

$$4\pi R_E^2 S = \pi R_E^2 S_0 \quad \text{or} \quad S = 340 \text{ W m}^{-2} \tag{2.7}$$

What is the temperature of the Earth's surface that establishes radiative equilibrium between the absorbed solar radiation $(S_0/4)(1-\alpha)$ and the outgoing terrestrial flux in the absence of an atmosphere?

$$\frac{S_0}{4}(1-\alpha) = \sigma T_e^4 \tag{2.8}$$

The planetary albedo α is 0.3, giving an *emission temperature* $T_e = 255$ K – rather cold. This is because we have neglected the atmosphere. If we include the atmosphere and model it as a single-layer black body at temperature T_A, we arrive at radiative balance conditions between absorbed and emitted radiative fluxes at the interface between Earth and space ("top of atmosphere" in climate science jargon), in the atmosphere, and at the surface (with temperature T_s):

$$\frac{S_0}{4}(1-\alpha) = \sigma T_A^4 \qquad \text{TOA} \tag{2.9}$$

$$\sigma T_s^4 = 2\sigma T_A^4 \qquad \text{atmosphere} \tag{2.10}$$

$$\frac{S_0}{4}(1-\alpha) + \sigma T_A^4 = \sigma T_s^4 \qquad \text{surface} \tag{2.11}$$

From which we find $T_A = T_e = 255$ K and $T_s = \sqrt[4]{2}T_e = 303$ K – somewhat warmer than the actual $T_s = 288$ K.

Figure: Hartmann (1994)

Radiative equilibrium

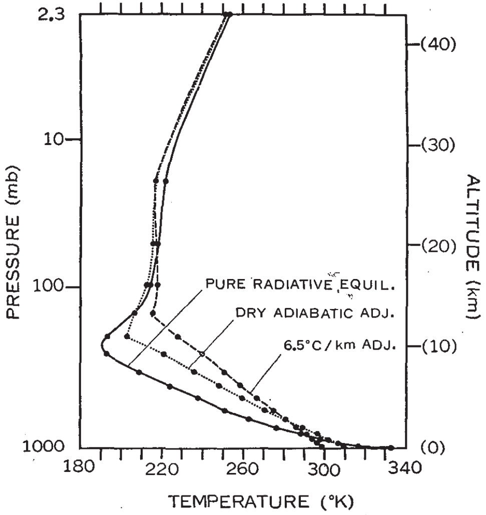


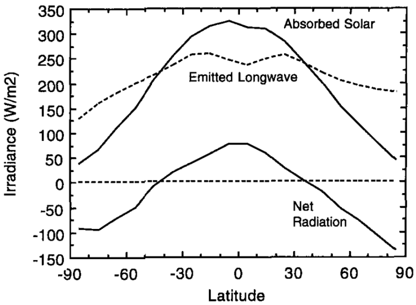
Figure: Manabe and Strickler (1964)

- Pure radiative equilibrium produces a very steep temperature gradient, requiring a high surface temperature
- The resulting atmosphere is not convectively stable
- Stability is restored through vertical latent + sensible heat transport that cools the lower layers and heats the upper layers of the troposphere
- The resulting atmosphere is in *radiative–convective* equilibrium
- (The actual 3D atmosphere is in radiative–convective–advective equilibrium)

Zonal-mean radiative balance at TOA

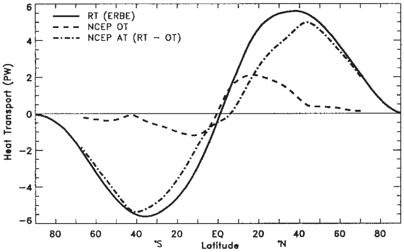
TOA radiative balance. . .

- The top-of-atmosphere (TOA) radiative balance measures how much energy enters or leaves the climate system
- In the tropics, the net energy flux is positive (into the climate system)
- In the extratropics, the net energy flux is negative (out of the climate system)



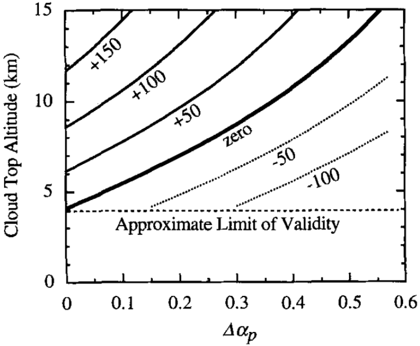
. . . requires meridional energy transport

- To maintain a steady state (e.g., constant long-term zonal-mean temperature), energy transport from the tropics into the extratropics is required
- The *meridional divergence* of the energy transport balances the radiative energy flux
- Contributions to transport from ocean and atmosphere – subject of the next section



Figures: Hartmann (1994), Trenberth and Caron (2001)

Clouds



Clouds have two effects:

- ≈ -50 W m^{-2} Increased reflection of shortwave ($\Delta\alpha$) $\rightarrow$ cooling
- $\approx +25$ W m^{-2} Colder longwave emission temperature $\rightarrow$ warming

so that the sign of the cloud contribution to the radiative balance depends on cloud top temperature and cloud thickness

- SCu Thick, low clouds (like stratocumulus) have a large cooling effect
- Ci Thin, high clouds (like cirrus) have a large warming effect

Figure: Hartmann 1994

Global-mean radiative energy fluxes

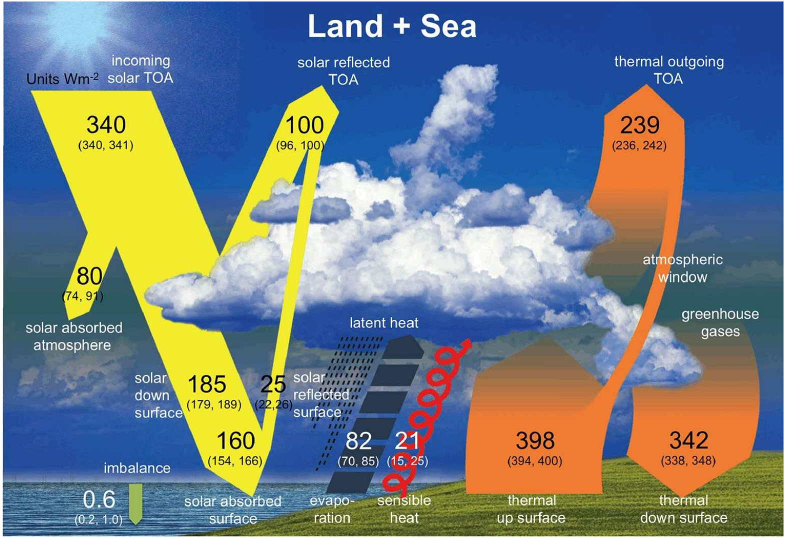


Figure: Wild et al., 2015