Lecture 24

Radiation and climate.

Objectives:

1. Global radiation budget.

2. Radiation in climate models.

Required reading:

L80: 2.3; 2.5; 8.3-8.5

Advanced reading

G&Y: 9

1. Global radiation budget.

- **Planetary radiative equilibrium**:

  TOA outgoing radiation = TOA incoming radiation

  (over the entire planet and long time interval, e.g., a year).

Let's estimate the **effective temperature** assuming that the Earth is in the radiative equilibrium.

The sun emits $F_s = 6.2 \times 10^{-7}$ W/m^2 (a blackbody with about T= 5800K).

From the energy conservation law, we have

$$\mathbf{F_s\, 4\pi R_s^2 = S\, 4\pi D_0^2}$$

where $\mathbf{R_s}$ is the radius of the sun (6.96×10^5 km);

$\mathbf{S}$ is the solar flux reaching the top of the atmosphere (called the **solar constant** = about 1368 W/m^2; see Lecture 4) at the average distance of the Earth from the sun, $\mathbf{D_0}$ = 1.5×10^8 km.

Thus we have

$$\boxed{\mathbf{S = F_s\, R_s^2 / D_0^2}} \qquad [24.1]$$

If the instantaneous distance from the Earth to sun is **D**, then the total sun energy flux $\mathbf{F_D}$ reaching the Earth is

$$\mathbf{F_D = S\,(D_0/D)^2} \qquad [24.2]$$

NOTE: Here $\mathbf{F_D}$ is the solar flux denoted by πF_0 in L80 and all previous lectures.

NOTE: the sun is a blackbody => $\mathbf{F_s = \pi I_s}$, where I_s is the sun intensity. The intensity (the energy per the solid angle) is invariant (i.e., intensity at the sun = intensity at the Earth). Because of the large distance, the solar radiation reaching the earth is plane-parallel (called collimated), so TOA solar intensity is $\mathbf{I_0=\pi F_0 \delta(\mu-\mu_0)\ \delta(\phi-\phi_0)}$

The total sun energy intercepted by the cross section of the Earth is $\mathbf{F_{s0}\ \pi R_e^2}$ where $\mathbf{R_e}$ is the radius of the Earth. This energy is spread uniformly over the entire planet (with surface area $\mathbf{4\pi R_e^2}$). Thus the amount of received energy per unit surface becomes

$$\mathbf{S\ \pi R_e^2 / 4\pi R_e^2 = S/4}$$

Therefore the total energy $\mathbf{Q_{in}}$ (in W/m^2) absorbed by the earth-atmosphere system is:

$$\mathbf{Q_{in} = (1 - r)\ S/4}$$

where ***r*** is the spherical (or global) albedo (see Lecture 18). Spherical albedo of the earth is about 0.3.

Assuming that the Earth is a blackbody with temperature T_e, we have:

$$\mathbf{Q_{out} = F_b = \sigma_B\ T_e^4}$$

where $\mathbf{\sigma_B}$ is the Stefan-Boltzmann constant.

From the balance of incoming and outgoing energy, the **effective temperature** of the Earth is:

$$\mathbf{Q_{in} = Q_{out}}$$

$$\mathbf{S\,(1 - r)/4 = \sigma_B\ T_e^4}$$

$$\mathbf{T_e^4 = S\,(1 - r) / 4\,\sigma_B}$$

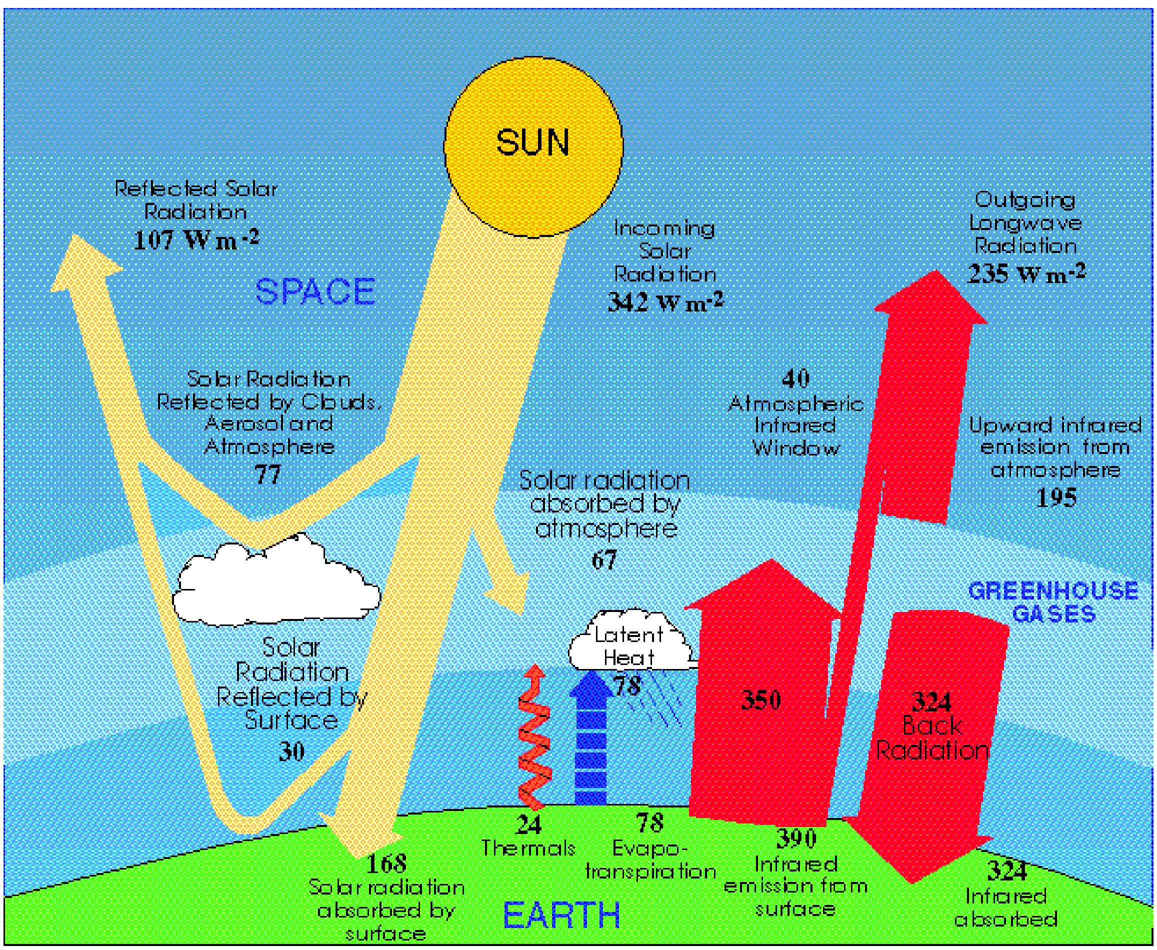
$$\mathbf{T_e = 255\ K = -18^0C\ is\ very\ low!!!}$$

Why? : because we didn't include the greenhouse effect and ignore the temperature structure.

Table 24.1 Effective temperatures of some planets in the radiative equilibrium.

Planet	Relative distance to the sun with respect to the Earth	Global albedo	T_e(K)
Mercury	0.39	0.06	441
Venus	0.72	0.78	226
Earth	1	0.3	255
Mars	1.52	0.17	217
Jupiter	5.2	0.45	106

Figure 24.1 Earth's energy budget (modified from Kiehl and Trenberth, 1997).



2. Radiation in climate models.

- Climate models may be classified by their dimensions:

Zero Dimensional Models (0-D):

consider the Earth as a whole (no change by latitude, longitude, or height)

One Dimensional Models (1-D):

allow for variation in one direction only (e.g., resolve the Earth into latitudinal zones or by height above the surface of the Earth)

Two Dimensional Models (2-D):

allow for variation in two directions at once (e.g., by latitude and by height)

Three Dimensional Models (3-D)

allow for variation in three directions at once (i.e., divide the earth-atmosphere system into the domains, each domain having its own independent set of values for each of the climate parameters used in the model.

- Climate models may be classified by the basic physical processes included into the consideration:

Energy Balance Models:

0-D or 1-D models (e.g., allow to change the albedo by latitude) calculate a balance between the incoming and outgoing radiation of the planet;

Radiative Convective Models:

1-D models to model the temperature profile the atmosphere by considering radiative and convective energy transport up through the atmosphere.

General Circulation Climate Models:

2-D (longitude-averaged) or 3-D models solve a series of equations that describe the movement of energy and momentum and the conservation of mass and water vapor; in general, GCMs use the two-stream approximation to solve the radiative transfer equation.

Underlying principles of a radiative convective model:

calculate upward and downward energy fluxes at levels in the atmosphere following a general scheme :

1.Set up a starting condition, usually an isothermal atmosphere.

2.Calculate the energy change in each layer of atmosphere resulting from an imbalance between the net radiation at the top and bottom of the layer.

3.Convert this to temperature change.

4.This leads to a new temperature profile which can then be compared with a predetermined lapse rate to see if the atmosphere is convectively unstable, if so mixing takes place before the process is repeated for the next time step.

5.Compare the temperature profile with one for the previous step and return to stage 2 if there is sufficient difference, otherwise end.

All radiative convective models have two boundary conditions :

1.At the top of the atmosphere the short wave and long wave energy fluxes must balance.

2.At the surface the net gain of energy by radiation equals the net loss by convection.