Lecture Ch. 12b

- Atmospheric heat engine
  - Latitudinal and meridional heat transfer
  - Walker circulation and Aus-Asia monsoons
  - Efficiency, irreversibility, entropy
  - Hydrological cycle
- Current research
  - Aerosols, precipitation, and evaporation

Curry and Webster, Ch. 12
For Tuesday: Read Ch. 13
For Dec 2: Review Past Homework, Quizzes, Reading, Midterm

The Atmospheric "Heat Engine"

Latitudinal variation in the net radiation flux at the top of the atmosphere results in an overall heat transport from equatorial to polar regions. In effect, the atmosphere operates as a heat engine, whereby a portion of the absorbed radiation (heat source) is converted into kinetic energy (work). The efficiency of the atmospheric heat engine is low, because of strong irreversibilities in the system arising primarily from a highly irreversible heat transfer of solar radiation to the Earth. Finally, the global hydrological cycle modulates the Earth's energy and entropy budgets through radiative and latent heating.

Atmospheric Heat Engine

The strength of the thermal circulation depends on the efficiency of the heat engine. For a reversible Carnot engine, we have from (2.30b)

$$\mathscr{E} = 1 - \frac{T_2}{T_1} = \frac{T_1 - T_2}{T_1}$$

If we identify $T_1 = 300$ K with the tropical surface heat source and $T_2 = 200$ K with the high-latitude upper atmosphere cold sink, we obtain $\mathscr{E} = 33\%$. Since the Earth's climate system is irreversible, the actual efficiency of the Earth as a heat engine is much smaller. A more meaningful estimate of the efficiency can be determined from (2.30a):

$$\mathscr{E} = \frac{w}{q_1}$$

The heating term is the mean incoming solar radiation, $q_1 = (1 - \alpha_P)S/4 = 238$ W m^{-2}. To estimate the work term, w, it is assumed that the production of kinetic energy is balanced by frictional dissipation, maintaining the average kinetic energy of the atmosphere. This term has been estimated by Oort and Peixoto (1983) to be $w = 2$ W m^{-2}, yielding an efficiency of $\mathscr{E} \sim 0.8\%$.

Atmospheric Entropy

- Difference between energy and entropy flux

The total flow of entropy at the top of the atmosphere is determined by the radiative transfer (Figure 12.9). Although the net incoming and outgoing radiation at the top of the atmosphere are equal when averaged globally and over an annual cycle (12.1), the net incoming and outgoing radiation entropies are never equal to each other. The solar radiation brings in a small amount of entropy in comparison with the entropy that longwave radiation removes from the system.

- Irreversible processes

implies the action of strong irreversible processes. Since the temperature of the Earth–atmosphere system is considerably lower than the sun, which is the source of the solar radiation, there is a highly irreversible heat transfer from the sun to the Earth. Scattering of radiation is an additional source of irreversibility.

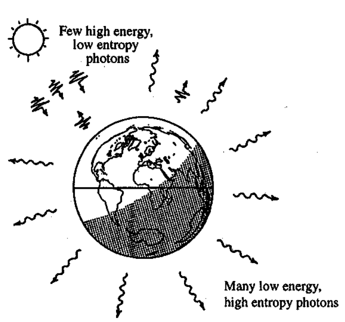

Figure 12.9 Radiative transfer determines the flow of entropy at the top of the atmosphere. Outgoing longwave radiation removes entropy from the system, while incoming solar radiation adds entropy. (After Stephens and O'Brien, 1993.)

Atmospheric Entropy

- Internal production of entropy by Earth

ated with the net solar radiation. This negative entropy stream at the top of the atmosphere allows internal production of entropy by the Earth while at the same time maintaining order in the atmosphere. If the overall entropy of the atmosphere were increasing, the atmosphere would approach a state of maximum entropy, leading to a uniformity of the climate.

Hydrological Cycle

- Definition

The continual movement of water among the reservoirs of ocean, atmosphere, and land is called the *hydrological cycle*. The total amount of water on Earth remains

- Residence times

the accumulation or depletion rate. The atmospheric residence time (mass of water vapor divided by total precipitation) is about 10 days; that is, the atmosphere recycles its water over 30 times per year. Surface water over land has a residence time of about 5 years, although the residence time for soil moisture is about 1 year and the total residence time for glaciers is 6000 years. The residence time of water in the oceans is about 3000 years, although not all parts of the ocean recycle at the same rate. In the ocean surface layers the time scales may be on the order of days to weeks, while deep bottom water may take thousands of years to recycle.

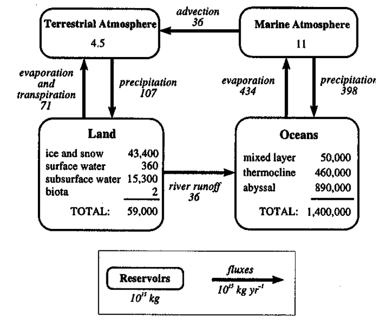

Figure 12.10 The global water cycle and the reservoirs of water. Arrows indicate fluxes of water from one reservoir to another. Note that water in the atmosphere over the ocean and the land accounts for only 0.0001% of the total water in the system. Units are in 10^{15} kg. (Data from Chahine, 1992.)

Table 12.1 Evaluation of the recycling rate (%), which is the amount of precipitation that comes from local evaporation versus horizontal transport, evaluated for scales of 500 and 1000 km. (Data from Trenberth, 1998.)

	scale (km)	
	500	1000
Global	9.6	16.8
Land	8.9	15.4
Ocean	9.9	17.3

Latitudinal Heating Distribution

- Net heating at equator
- Net cooling at poles

Figure 12.4 shows that the annual mean net radiation is positive equatorward of 40° latitude and negative at higher latitudes. Since polar temperatures are not observed to cool and tropical temperatures are not observed to warm on average, a transport of heat from equatorial to polar regions must occur. This transport occurs via fluid motions in the atmosphere and ocean that are driven by horizontal pressure gradients generated by the uneven heating.

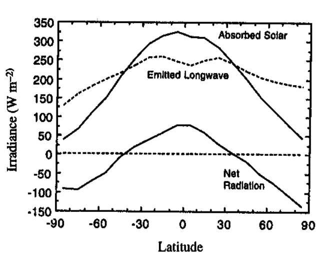

Figure 12.4 Annual mean absorbed shortwave, outgoing longwave, and net radiation averaged around latitude circles. (From Hartmann, 1994.)

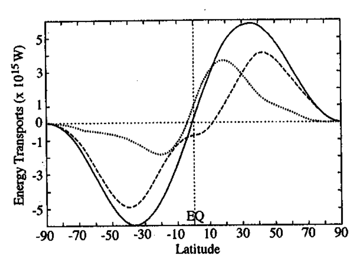

Figure 12.5 Annual mean northward energy transports required to equalize the pole–equator radiative imbalance. The solid line represents the top-of-the-atmosphere radiation budget, the dashed line represents the atmosphere, and the dotted line represents the ocean (From Zhang and Rossow, 1997).

Heating and Circulation

- Fluid motion from vertical density gradient

Suppose an initially barotropic atmosphere is heated at low latitudes and cooled at high latitudes in a manner such that there is no net heating over the globe. In a hydrostatic atmosphere, the thickness of a layer between isobaric surfaces (1.45) increases at low latitudes and decreases at high latitudes, tilting the isobaric surfaces. This produces a nonuniform distribution of density and temperature on isobaric surfaces (a *baroclinic atmosphere*), and hence a horizontal pressure gradient that results in potential energy being available for conversion to kinetic energy.

This process is illustrated in Figure 12.6 by two immiscible fluids of different densities that are adjacent to each other. Assuming that both fluids are in hydrostatic equilibrium, a pressure gradient force is directed from the heavier fluid to the lighter one, causing the heavier fluid to accelerate towards the lighter one. The ensuing motion will result in the heavier fluid lying beneath the lighter one. Through the sinking of denser fluid and the rising of the lighter fluid, the center of gravity of the system is lowered and potential energy is converted into kinetic energy of fluid motions.

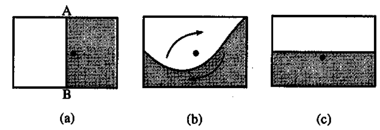

Figure 12.6 (a) Heavier (shaded) and lighter fluids separated by a movable partition, AB. The dot represents the center of gravity. (b) Fluids in motion following the removal of the partition. (c) Equilibrium configuration of the fluids after the motion has dissipated.

Meridional Heat Transfer

- Equator-to-pole heat transport

If the Earth were not rotating, the atmospheric transport of heat from pole to equator would occur as a direct thermal circulation: heating at the surface in the equatorial regions causes rising motion → heat is transported polewards at upper levels → sinking occurs over the polar regions → the circulation is completed by a low-level return flow of cold air from high to low latitudes. The actual mean equator-to-pole transport of heat in the atmosphere is complicated considerably by the Earth's rotation, angular momentum considerations and subsequent hydrodynamical instabilities, especially poleward of the subtropics. The large-scale eddies (e.g., storms) produced in midlatitudes rapidly transfer heat poleward to satisfy the global energy balance.

Zonal Heat Transfer

- Walker circulation
- Asian-Australian monsoon

In addition to the global meridional transfer of heat from low to high latitudes, heat transfer occurs on large horizontal scales, primarily in response to turbulent heat fluxes into the atmosphere arising from surface temperature gradients arising from the geographical distribution of continents. The *Walker Circulation* (Figure 12.7) is generally symmetric about the equator with ascending motion in the warm pool regions of the Indian and Pacific Oceans and the Indonesian rchipelago, and descent in the western Indian Ocean and the eastern Pacific Ocean. Weakening or reversal of the Walker circulation, where there is rising motion in the eastern Pacific and sinking motion in the western Pacific, occurs several times in a decade and is referred to as *El Niño*. The Asian–Australian *monsoon* (Figure 12.8) is a global circulation pattern which is asymmetric about the equator and has its focus and basic forcing in the land/ocean distribution of the Eastern Hemisphere. If there were no tropical continents and the

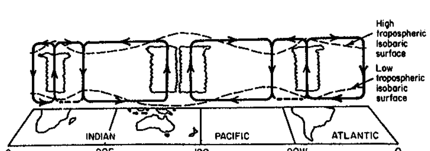

Figure 12.7 Schematic view of the east–west Walker circulation along the equator, indicating low-level convergence in regions of convection where mean upward motion occurs. (From Webster, 1987. © John Wiley & Sons, Inc. Reprinted with permission.)

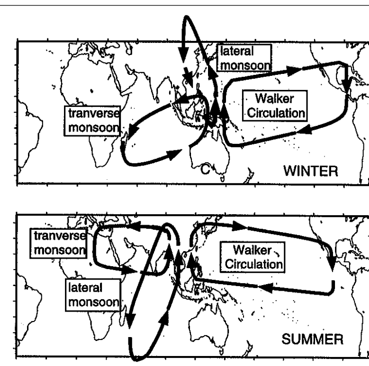

Figure 12.8 The Asian–Australian monsoon circulation pattern. The somewhat complicated wind circulation patterns and the seasonal differences in the patterns result from the variations in the heating of the continents. (From Webster *et al.*, 1998.)