Introductions

Jargon: be sure to tell me if you don't know the meaning of a word that I am using.

In this class, we will be interested in the role that the ocean plays in the climate system, particularly under global warming. To do that, we first have to understand how the ocean and atmosphere interact in an equilibrium climate, and what variability exists in the natural climate system. Then we will begin to investigate how what role the ocean plays in the climate system under global warming.

During the past century we have seen an increase in temperature of a little over 0.5 °C. The century scale upward trend is now generally accepted to be the consequence of increasing CO2 in the atmosphere. However, superimposed on this warming trend are other signals of shorter period climate variability that have interannual and interdecadal times scales that may involve involve reorganizations of the ocean and atmosphere circulation. They are also associated important changes in oceanic ecosystems and regional weather patterns. In order to understand the climate system, we must focus also on these patterns of variability.

Energy balance of the planet

To understand the climate system, we must understand the energy balance of the planet.

Simplest example: Imagine a planet without fluid (either atmosphere or ocean). What happens?

The sun can be considered as a black body that obeys the Stefan Boltzman Law. That Law states that the energy flux per unit area per unit time for a block body obeys tha equation

$$E = \sigma T^4 \tag{0.1}$$

where

$\sigma = 5.67 \times 10^{-8} Wm^{-2} K^{-4}$

The energy flux is thus sensitive to temperature T. The sun has a temperature of approximately 6000K and radiates primarily in the

visible part of the spectrum, this radiation is termed short wave radiation.
Thus, the total energy radiated by the sun every second will be given by (1.1) multiplied by the area of the sun or

$$I_0 = \sigma T^4 \times 4\pi R^2{}_{sun} = 3.9\times 10^{26} W = 3.9\times 10^{26} J/s$$

Here R_{sun} is the radius of the sun.

Now we want to know what the amount of energy that hits earth from the sun. That will be given by

$$Q_0 = \frac{I_0}{4\pi R_{SE}^2} = 1367 W/m^2$$

where R_{SE} is the distance from the earth to the sun. To put this in perspective, over square meter, this is about the energy used by a hairdryer. This would be the insolation seen by someone in the Sahara dessert at about noon.

The daily averaged insolation will be less because the earth rotates once a day and it is not a flat disk, rather it is a sphere. The average daily insolation that reaches the planet is thus given by

$$Q_0/4 = Q_0 \times \frac{\pi R_E}{4\pi R_E^2}$$

The most of the solar radiation is absorbed by the earth and then is reradiated at a lower temperature and longer wavelength. To determine the average temperature of the planet, we note that the incoming solar radiation must be balanced by reflected outgoing radition plus radiation from the planet. In equation form this looks like

$$\frac{Q_0}{4} - \alpha\frac{Q_0}{4} = \sigma T_E{}^2$$

where α is the albedo of the planet. The global average is about 0.3. Over the ocean however, on average, the albedo is about 0.05. We can solve for the temperature of the earth T_E.

$$\frac{Q_0}{4}(1-\alpha) = \sigma T_E{}^4$$

$$T_E = \left(\frac{Q_0}{4\sigma}(1-\alpha)\right)^{1/4} = 255K = -18C$$

The average temperature at the surface of the planet is about 15C or 288K. Surface warming owing to the greenhouse effect warms the planet. We will come back to this later.

Note that the magnitude of the longwave radiation up is about 240 W/m^2

Surface heat budget over the oceans

We will examine the surface heat budget over the ocean in more next.

In an annual averaged sense, the ocean is heated by incoming solar radiation and downward longwave radiation from the atmosphere, while it is in general cooled by longwave radiation and both sensible and latent heat flux. The ocean's albedo is on average about 4-5% which is much less than the global average. In equation form, we have

$$Q_{solar} + Q_{LWdown} = \alpha Q_s + Q_{LWup} + Q_{H(sensible)} + Q_{E(evaporative,latent)}$$

We will look at the the turbulent fluxes of heat, sensible heating which is physical movement of heat from the ocean to the atmosphere, while the latent heat flux is the flux of heat from the ocean that is released when sea-water is evaporated.

Sensible heat flux comes about when there is shear driven turbulence that comes from the interaction of the wind with the ocean surface. The heat flux
We can write the sensible heat flux with the formula

$$Q_H = c_p \rho_A \overline{w'T'}$$

Here c_p is the specific heat at constant pressure of air while ρ_A is the atmospheric density (about 1.4 kg/m3).

Likewise for sensible heat flux we have

$$Q_E = L_V \rho_A \overline{w'q'}$$

where L_V is the latent heat of vaporization.

Here w' is the vertical velocity of turbulent eddies, while T' is the temperature carried by that vertical velocity. Notice that if an eddy is carrying air parcels upward, and the temperature is warmer than its surroundings then multiplying the two together gives a positive heat flux, while if vertical velocity is downward and the temperature is lower than its surroundings, then the heat flux will still be upwards.

Since the vertical velocity and perturbations of humidity and temperature are difficult to measure, we instead establish what we call a bulk formulae that is based on the fields that are measurable, such as temperature or humidity averaged over many eddies. What we use instead is

$$Q_H = c_p \rho_A U_r C_{DH} (T_S - T_A(z_r))$$
$$Q_E = L_V \rho_A U_r C_{DE} (q_S - q_A(z_r))$$

Here z_r is a reference level for measurements of the atmospheric variables, and q_s is the saturated humidity value at the sea surface temperature. Typically z_r is 10m. Each of the turbulent fluxes depends on a drag coefficient that is unitless.

$C_E = 1.5 \times 10^{-3}$

$C_H = 1 \times 10^{-3}$

In general the sea surface is warmer than the atmsphere over head (although not always) and the sensible and latent heat flux results in heat taken from the ocean and transferred into the atmosphere.

Latent heating tends to be larger in the tropics and while sensible heating is larger at high latitudes where the atmosphere is cold.

Equation of state of sea water

For the purpose of this class, we need to know that the density of sea water depends on temperature, salinity and pressure and that there is a lot of structure in the oceanic density field at the 0.1% level. The equation of state is non-linear in temperature and salinity and has the following properties

1. At high T, density is more sensitive to T than S
2. At low T, density is more sensitive to S
3. The maximum density of fresh water occurs at 4C while the maximum density of sea water is at its freezing point. This changes the properties of overturning in the ocean verses a fresh water lake.
4. The freezing point of seawater is about -1.9C.
5. Sea water is compressible, in situ density varies by a few %
6. The minimum density of sea water is about 1021 kg/m3, while the maximum is about 1071 kg/m3

We defined a parameter σ

$\sigma = \rho - 1000$

where ρ is the density of sea water.

σ is a function of temperature, salinity and pressure

$\sigma = \sigma(T, S, p)$
We also define
$\sigma_t = \sigma(T, S, 0)$
and the potential density
$\sigma_\theta = \sigma(\theta, S, 0)$
such that the potential density is the density that sea-water would have if brought to the surface adiabatically (without exchange of heat with its surroundings). It is typically less than in situ density because of the compressibility effects. When calculating geostrophic currents, we have to use the potential density.