In this section of the course, we'll talk about the incredibly important role that the oceans play in the climate system. We'll begin by reviewing some of the most important considerations. First, the ocean has almost all of the water on the planet. It has an immense heat capacity compared to the atmosphere, and we'll explore the role that that plays in the climate system. It retains a memory of past disturbances that can extend to thousands of years. It exchanges energy and moisture with the atmosphere and transports energy in very large amounts. It absorbs, stores, and ejects carbon dioxide in very large amounts. It is the site of a large fraction of the biological activity on earth, and it is a major component the biogeochemical cycles that are so important for the evolution of climate on long time scales. When it is frozen, it can undergo very large albedo changes because the albedo of sea ice is very much larger than the albedo of sea water.

Let's compare the ocean to the atmosphere which we talked about in the previous section of the course. We'll begin by talking about similarities to the atmosphere. It turns out the governing equations, particularly the equations of motion, are nearly identical to those of the atmosphere. And like the atmosphere, it is a global-scale fluid on a rotating earth. Major differences from the atmosphere include the fact that the ocean has continental barriers to east-west motions. We don't see analogs to this in the atmosphere. North-south mountain ranges provide a partial barrier, but not a complete barrier, to east-west motion. We'll see that this plays a very important role in the circulation and transport properties of the ocean.

The oceans are virtually opaque to radiation at all wavelengths, both solar and terrestrial. Almost all of the solar radiation is absorbed in the first 100 meters or so of the ocean, for example. And the ocean is both heated and cooled very close to its upper surface. This is very different from the atmosphere which has internal heat sources and sinks, and makes the ocean a very inefficient heat engine.

There is no equivalent in the ocean to moist convection, although the density of seawater depends on pressure, temperature, and salinity, and variations of salinity introduce issues that are somewhat analogous to the condensation and evaporation of water in the atmosphere. Finally, the oceans are much more difficult to observe than the atmosphere, thanks to the fact that frequencies we use to transmit information by radio are very strongly absorbed in the ocean. So it's not possible to send probes into the ocean the way we do in the atmosphere and have those probes radio information back to the surface.

So how do we observe the oceans? As we've just stated, radio waves don't travel through seawater. It's not possible to relay information from sensors. Therefore, we have to make in situ measurements to understand what's happening below the surface of the ocean. Historically, these have been very expensive because they rely on ships. So here, for example, are two pictures of ships that have been used historically by the Woods Hole Oceanographic Institute. The sailing vessel that you see here on the left was used in the early and middle 20th century, more recently been replaced by a modern research ship that you see on the right. These are very effective platforms but very expensive to operate.

How does one actually make measurements at depth? Well, there are various ways to do this. Here is one example called a Niskin bottle. This instrument is lowered over the side of the ship to various different depths. A bungee cord is used to open the ends of the cylinder, which then fill with water. The end are then sealed, and the instrument is brought back to the surface where the seawater is analyzed. This is obviously an expensive and labor-intensive enterprise.

More recently, the study of the oceans have greatly benefited from the advance in robotics. Here is a picture, on the left, of an instrument called an Argo float, which is initially launched from a ship and then automatically descends, makes measurements, comes back to the surface and transmits the information to satellite. It then goes back down to make more measurements, comes back up and so forth. The map at the right shows the positions of these Argo floats, as of the middle of December in 2013. These floats just drift around with ocean currents. They have to be occasionally re-deployed, but they're a very effective new means of sampling the ocean.

Let's now talk about the role of the ocean as a thermal reservoir. We'll start by noting that the mass of the ocean is about 350 times larger than the mass of the atmosphere and has a heat capacity about 500 times the effective heat capacity of the atmosphere. In defining the latter, we take into account the latent heat conversions that also occur in the atmosphere. But any way you measure it, the ocean has an enormous heat capacity relative to the atmosphere.

We observe that heat is mixed very rapidly to the top of the ocean, the so-called upper ocean mixed layer, which is kept turbulent by a combination of winds and convection. This turbulent layer is anywhere from about 20 to about 150 meters in depth, depending on location and season. Heat is very rapidly mixed through this layer, but may take a much longer time to get into the deep ocean.

Well, let's just talk about the interaction between the ocean's mixed layer and the atmosphere. And we'll represent that interaction by this very simple equation. So on the left hand side, we have essentially the time rate of change of the upper ocean's heat content. So C sub l is the specific heat capacity of seawater, rho sub l is its density, and h here is the depth of the upper ocean mixed layer. These are multiplied by the time rate of change of the ocean mixed layer temperature, dT by dt. That is being driven by two terms on the right. The first term is the net radiative flux into the ocean from the atmosphere, F sub rad, and the second term which is negative is the turbulent flux of heat from the ocean to the atmosphere. It's represented by dimensionless exchange coefficient times the density of air times the near surface wind speed times the specific heat capacity at constant pressure of air times the difference between the ocean temperature and the atmospheric temperature.

Now we have neglected quite a few terms in developing this equation. We've neglected the latent heat flux from the ocean to the atmosphere, which in practice is the bulk of it, particularly in the tropics. We've neglected any flux of heat between the ocean's mixed layer and the deep ocean. And we've neglected horizontal advection of heat. But we're using this equation not to make exact predictions, but to make a narrow point about the behavior of the system. The various terms in this equation are defined on this slide.

Now we can rewrite that equation in this form, by taking a special circumstance in which the radiative flux does not vary in time. And we assume that the atmospheric temperature, T sub a, is just white noise, that is noise whose amplitude does not change over the span of frequencies represented in the atmosphere. Then we can rewrite the equation on the previous slide this way, time rate of change of the upper ocean temperature here on the left is proportional to a term q minus T divided by tau, where q is just a white noise term, is just varying randomly in time. And tau is defined here, according to the constants in the equation on the previous page. If we plug-in typical numbers for these constants, tau has a time scale of roughly a year. And we're going to solve this equation by taking q to be randomly varying, every three days, with a zero mean.

We'll simply solve that numerically, which is very easy to do. And here is one particular solution of that equation, marching forward in time from a particular initial condition over a period of 20 years. And what's graphed here is the ocean mixed layer temperature perturbation from its initial value, in tenths of a degree. And one can see that there are all kinds of interesting time scales of variations, not only variations on the time scale in which the forcing is varying, which is only three days, hardly detectable on this graph. But one can see that the response is considerably redder than the white noise we're using to drive it. That is, we see variations on time scales of years and even decades in this response, showing that the ocean takes the white noise of the atmosphere and makes it very red. The ocean reddens the noisy input.

And this is an interesting if fairly simple aspect of the ocean. Thanks to its enormous heat capacity, it can easily fool us into thinking that there are actually modes of oscillation of the ocean on time scales of years or decades. Now such modes undoubtedly exist, but it's easy to be fooled from something simple like a time series of ocean temperature. One cannot make the inference from a graph like this that there are actually modes of oscillation of the ocean on these time scales [such variability may just be random, rather than periodic].