Up until now, we've talked about the physics of various components of the climate system, and how they interact to produce the observed climate of the Earth. This puts us in a position to understand how the Earth's climate responds to various changes. Here we're going to introduce the concept of climate forcings, feedbacks, and the notion of climate sensitivity.

This diagram, which we've seen once before in this course, is an attempt to illustrate all the various processes that affect the Earth's climate. We can talk about changes in the climate that are forced purely externally-- that is, by things that don't themselves respond to climate change. Perhaps the cleanest example is changes in the sun, which we'll talk about in this section. Changes in solar forcing clearly can affect climate.

Another example of an external forcing is volcanic activity. Volcanoes put large amounts of material in the atmosphere, which can interact both with sunlight and with infrared radiation to cause climate changes. Volcanoes are not thought to react, on the other hand, to climate change. So they can be considered to be an external forcing.

In addition to forcings, there are feedbacks. For example, clouds respond quite quickly to changing conditions in the atmosphere. By interacting with both solar and infrared radiation, we've seen that clouds can have a profound influence on climate. So if one changes the climate by some external means, such as by changing sunlight, clouds will presumably react indirectly to those changes and affect the climate. That process is called a feedback.

Another example of a feedback is sea ice. Here, if we change the temperature of the oceans, or the wind circulation across the surface of the ocean, this may change the sea ice content. But sea ice has a very high albedo. That is, it reflects quite a bit of sunlight. Changing sea ice, therefore, can affect the climate. That's another example of a feedback.

Now, when we go to the composition of the atmosphere, whether we regard changes as forcings or feedbacks depends upon the time scale. Take a pretty clear example-- water vapor, H_2O, the most important greenhouse gas, has a lifetime in the atmosphere of about two weeks-- that is, it is constantly cycling through the atmosphere.

And the water vapor concentration is controlled mostly by temperature. So changes in the external forcing of the system-- for example, by changing sunlight-- will change the temperature of the atmosphere, which will change the water vapor content, which in turn will feed back onto climate. So for time scales in excess of a few weeks-- that is, almost all the time scales that we consider to be part of the climate system-- we have to regard water vapor as a feedback in the system.

But to take an example at the opposite extreme, carbon dioxide has a lifetime in the atmosphere of from hundreds to thousands of years. So if we're worried about time scales of 100 or 200 years, we regard carbon dioxide mostly as a forcing. That is, if we change the carbon dioxide content, it takes a long time for natural processes to alter the carbon dioxide content again. So we tend to regard carbon dioxide as a forcing on time scales of centuries or less.

This diagram shows many other kinds of feedbacks in the climate system, including biological feedbacks. The biosphere responds to climate change by changing the density and type of vegetation, for example. And yet on short time scales, we can regard some aspects of this as a forcing. So for example, changes in the use of land by human beings can temporarily eliminate forests, or change the character of the land, and thereby alter climate on time scales less than the time scale on which the biosphere recovers, or goes back to its native state. We might consider that a forcing.

This is a fairly comprehensive diagram coming from the last report of the IPCC, showing all kinds of different forcings and feedbacks in the system. Beyond just tallying what those forcings and feedbacks are, let's try to understand the concept a little bit more quantitatively. And to do that, we're going to begin with a very simple equation for the energy balance at the top of the atmosphere.

So one considers this to be the energy balance averaged over a year and integrated over the whole planet. That energy balance says that the net flux at the top of the atmosphere is equal to the net downward solar flux minus the net upward infrared flux. If the planet is in thermal equilibrium, these two fluxes balance, and the net top-of-the-atmosphere flux vanishes.

This net top-of-the-atmosphere flux may be regarded as a function of the surface temperature, T sub S, and many other variables x. These could be variables like carbon dioxide, surface albedo, and so forth. So let's just write this symbolically as the net flux at the top of the atmosphere being a function of surface temperature, T sub S, And then n variables, which we'll just call x-- x sub 1, x sub 2, and so forth.

Now, if we apply the chain rule to this, this says that an increment in the top-of-the-atmosphere energy flux is equal to the partial derivative of that flux with respect to the surface temperature, times an increment of surface temperature, plus the sum over n of the partial derivatives of the top-of-the-atmosphere energy flux with respect to the x's, times increments in the x's.

Now, we're going to regard these changes as slow enough that the net top-of-the-atmosphere radiative flux actually remains zero. So if we're always in equilibrium, then the sum of the terms on the right-hand side of this equation here must be 0.

Now let's just arbitrarily call the last x, the n-th value of x, a forcing. We'll just give it a special name, delta Q. And it's special in the sense that we're going to externally specify what this forcing is. Let's think of it for the moment as a change in sunlight.

So we'll simply rewrite that chain rule by separating out that special n-th process. So the change in the top-of-the-atmosphere net radiation-- which we'll always force to be 0 in equilibrium-- is the derivative of that radiation with respect to surface temperature, times an increment of surface temperature, plus a sum to the n minus 1-th process of the change of net radiation with respect to x times the delta x's, plus this last 1 delta Q, which we'll regard as the forcing.

Now, we'll do one last operation on this equation, which is to recognize that the x's themselves can be regarded as a function of surface temperature, so the delta x can be written as delta x sub i with respect to surface temperature, times an increment of the surface temperature.

Now, if we divide this entire equation through by delta Q, and recognize that the left-hand side is 0, we get an expression for the rate of change of surface temperature with respect to the forcing-- d T sub s d Q. We're going to give that a special symbol, lambda R, which we'll call the climate sensitivity.

Lambda R is the climate sensitivity. And it's equal to minus 1 divided by the rate of change of the net radiation with respect to surface temperature, plus the sum to n minus 1 of the rate of change of net radiation with respect to the variables x, multiplied by the partial derivative of the x's with respect to surface temperature.

Now we'll define another quantity, uppercase S, as the inverse of minus the rate of change of the top-of-the-atmosphere radiation with surface temperature. That is the sensitivity of the climate that would occur without feedback. So this is if we eliminate the x's and we just have the Planck function feedback, partial TS with respect to changes in top-of-the-atmosphere radiation is called the climate sensitivity. It tells us how much the surface temperature changes for a given change in the external forcing.

So with lambda R called the climate sensitivity, and S defined this way, we can rewrite the equation on the previous slide this way. So the climate sensitivity is equal to the climate sensitivity S without feedbacks, divided by 1 minus S times the sum of these terms. And it has to be recognized that these terms can be of either sign.

When these terms are positive, they're called positive feedbacks. When they're negative, they're called negative feedbacks. But it's very important to understand from this equation that the feedback factors do not add linearly in their collective effects on sensitivity. That is, if we have one positive feedback factor and we add another positive feedback factor to that, the cumulative effect on the net climate sensitivity is somewhat greater than if you were just to add those linearly. And that's because of the form of this equation.

Notice also that if you have too many positive feedbacks or they're too strong, the denominator could potentially be zero or less. And then the equation doesn't make any sense. That corresponds to something called runaway feedback, where the positive feedbacks are so strong that the system simply cannot reach an equilibrium.