4 How Climate Changes

4.1 Forcing, Response, and Sensitivity

- FORCING (Watts per sq meter) is the difference between HEAT IN and HEAT OUT
- RESPONSE (degrees) is the change in global mean surface temperature required to re-establish thermal equilibrium (compensate for climate forcing)
- SENSITIVITY (degrees per W m^{-2}) is the ratio of RESPONSE to FORCING
- BASELINE sensitivity ~ 0.27 K W m-2 is what would happen without FEEDBACK

4.2 Climate Feedback

4.2.1 Positive Feedback Processes Amplify Climate Response

Water vapor feedback

Ice/Albedo Feedback

High Cloud Feedback

4.2.2 Negative Feedback Processes Reduces Climate Response

Vertical Mixing feedback

Low Cloud Feedback

4.3 Estimating Climate Sensitivity Including all the Feedbacks

- Estimating Sensitivity from Past Climate Change
- Estimating Sensitivity from Physics and Models

4.4 Climate Forcing by CO_2 and other Greenhouse Gases (GHGs)

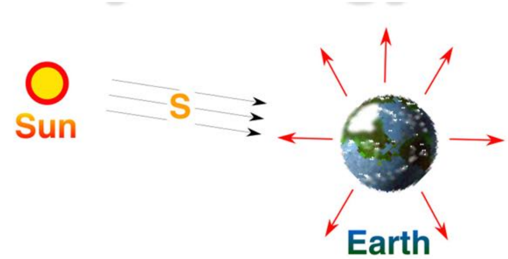
4.4.1 Ridiculous Eyewear Fashions

4.4.2 CO_2 Acts Just Like Sunglasses for OLR

4.4.3 Equilibrium Warming by CO_2

4.5 Comparing Climate Forcing, Response, and Sensitivity

4.1 Forcing, Response, and Sensitivity



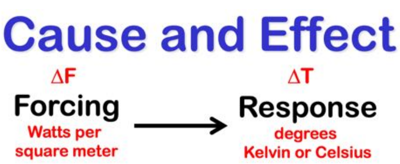
In Module 3, we considered how climate works when the flows of radiant energy into and out of the planet are *balanced* and global temperature is constant. In this module we consider what happens when energy is *unbalanced* – that is, *how global temperature can change*.

We define ***radiative forcing of climate*** as

$$\Delta F \equiv F_{in} - F_{out} \tag{4-1}$$

where F_{in} and F_{out} are (as defined in Module 3) the rates of energy flow into and out of the planet in Watts per square meter. When ΔF is positive, Earth progressively accumulates heat energy and the global mean temperature rises. When ΔF is negative, Earth progressively loses heat energy and the global mean temperature falls.

By the laws of thermodynamics a sustained imbalance of radiation flows into and out of the planet is the only way the average temperature of a planet can change!



We define the ***response ΔT_{eq} to climate forcing as the required change in global average surface temperature to re-establish thermal equilibrium*** and compensate for the radiative forcing.

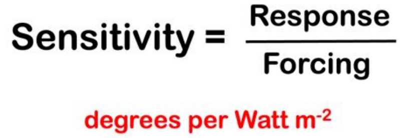
Climate response to radiative forcing has units of degrees Kelvin (of Celsius).

Since any radiative forcing will produce a climate response which takes some time to equilibrate, we refer to ***instantaneous forcing as the difference in radiation flows holding temperature constant***.

The ratio of climate response (in Kelvin) to radiative forcing is called ***climate sensitivity***

$$S = \frac{\Delta T_{eq}}{\Delta F} \tag{4-2}$$

and has the units Kelvin per (W m^{-2}). We can think of climate sensitivity as being analogous to a heat capacity on the scale of a planet: it's the amount of warming or cooling that will result from a radiation imbalance of a given strength.



We can derive a simple estimate of climate sensitivity that apples to the simple situation of a small rock in space from the Stefan-Boltzmann Law:

$$F = \sigma T^4$$

Taking the first derivative of the flux F with respect to temperature T, we get

$$\frac{dF}{dT} = 4\sigma T^3$$

Rearranging to isolate the change in equilibrium temperature

$$\Delta T_{eq} = \Delta F / (4\sigma T_{eq}^{3}).$$

Substituting Earth's current equilibrium temperature of 255 K, we calculate

$$S_{BB} = \frac{\Delta T_{eq}}{\Delta F} = \frac{1}{4\sigma T^3} = 0.266\ \frac{K}{W\ m^{-2}} \tag{4-3}$$

We call this the ***baseline climate sensitivity***. The baseline climate sensitivity is the change in temperature of any blackbody in response to a radiation imbalance of 1 W m^{-2}.

The simple interpretation is that if everything else stays the same, each change of the net radiation (in minus out) will produce a change in equilibrium temperature of about a quarter of a degree Kelvin (or Celsius since Celsius and Kelvin degrees are the same size).

We can generalize the concept of climate sensitivity to apply to the global average surface air temperature Ts rather than just the radiative equilibrium temperature Teq. Recall from Module 3 that Earth's average surface temperature is much warmer than the radiative equilibrium temperature which instead applies relatively high up in the atmosphere where it's much colder.

Here we're treating the rest of the climate system as if it were a "black box" which somehow processes the radiative equilibrium and produces a surface climate (Fig 4-1). The real world isn't a mysterious black box but rather consists of pretty well-understood processes like clouds, oceans, land, thermals, greenhouse gases, and so forth. But we don't need to quantify all those complex processes to assess climate sensitivity. We just need to measure changes in surface climate in response to changes in radiative forcing.

Now suppose we tweak the radiation at the top of the atmosphere by 1 W m^{-2} either by adding solar radiation or reducing OLR or some combination. Climate sensitivity is assessed by asking the question: given an energy imbalance of 1 W m^{-2}, by how much will the average surface temperature change?

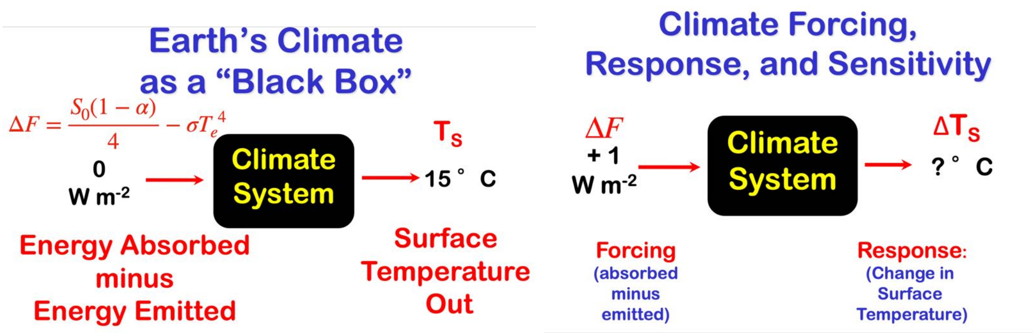


Figure 4-1: (Left panel) Earth's climate system as a "black box" that somehow converts radiation at the top of the atmosphere into surface temperature. In this example with thermal equilibrium (zero climate forcing), the average surface temperature is 15 ℃. (Right panel) Climate sensitivity is the ***change*** *in average surface temperature ΔT_S given an imbalance $\Delta F = 1$ W m^{-2} at the top of the atmosphere.*

4.2 Climate Feedback

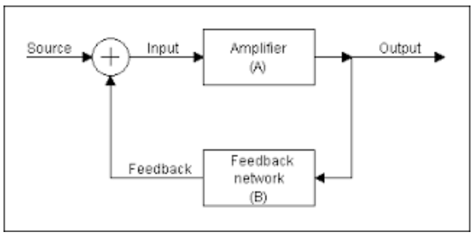
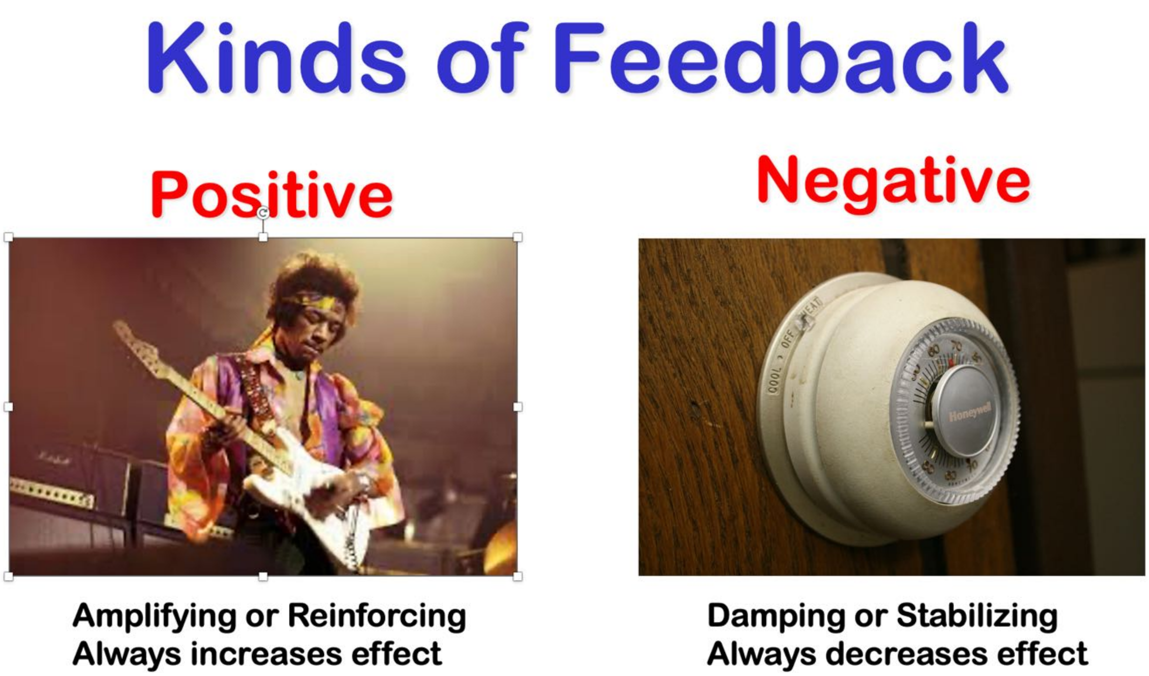


Figure 4-2: Feedback in an electrical circuit

But Earth is not just a rock in space!

Many aspects of Earth's climate interact with its surface temperature. For example, when Earth's surface (mostly ocean) warms, water vapor is transferred to the atmosphere, increasing the strength of the Greenhouse Effect. When the surface cools, snow and ice may accumulate and increase planetary albedo. Each of these changes will in turn cause further changes (either increases or decreases) to the surface temperature which can then cause more changes in the variables that determine climate.



These interactions are called climate feedback processes and they make Earth's climate more complicated and beautiful and worthy of a lifetime of study than that of a lump of rock in space!

We define ***positive feedback as anything that amplifies (increases) the climate response***. Conversely ***negative feedback is anything that damps (decreases) the climate response***.

CAUTION: We're using the word feedback in a formal way that comes from engineering and NOT in the colloquial (slang) American English sense! In everyday usage "positive feedback" pretty much always means something good (like a pat on the back or a thumbs-up or a big hug) and negative feedback means something bad (like a frown or a shaking head or a thumbs-down).

Climate feedback is neither "good" nor "bad." Positive feedback makes climate change more and negative feedback makes climate change less. If forcing is positive and climate is warming, then positive feedback makes it warm more. If forcing is negative and climate is cooling, positive feedback makes it cool more.

NOTE: It is **NOT TRUE** that positive feedback always implies warming or that negative feedback always implies cooling!!

Total climate sensitivity always includes all the feedbacks, positive and negative. If positive feedbacks are stronger than negative feedbacks, overall climate sensitivity will be greater than the baseline (blackbody) climate sensitivity. Conversely if negative feedbacks are stronger than positive feedbacks, overall climate sensitivity will be less than the baseline (blackbody) climate sensitivity.

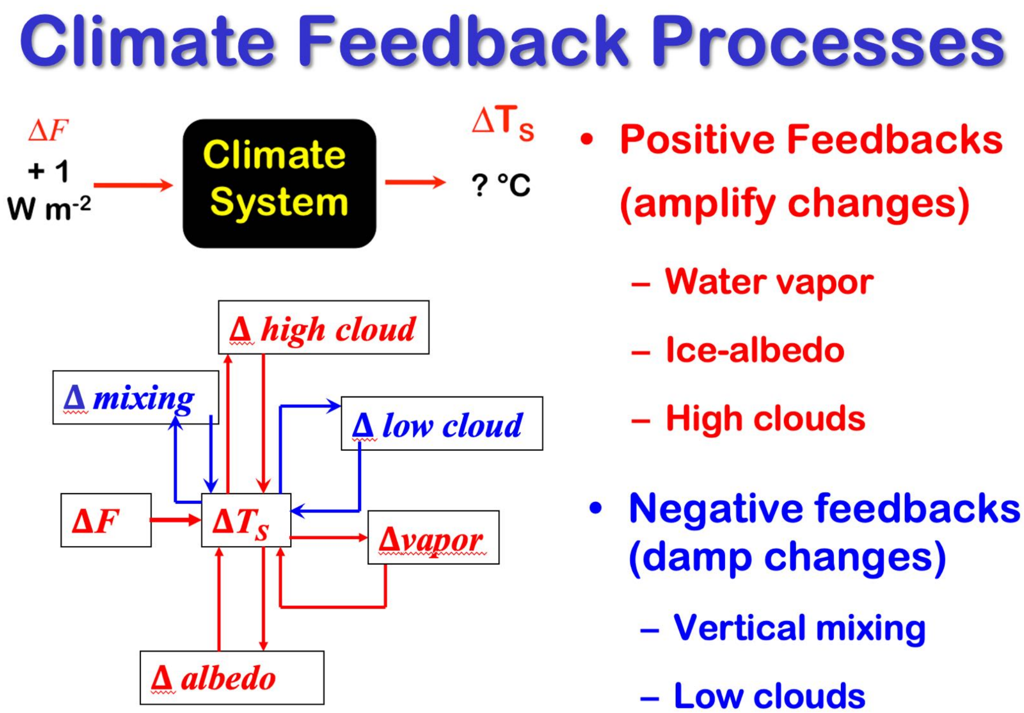


Figure 4-3: Climate feedback processes ***amplify*** *(positive feedbacks in red) or* ***reduce*** *(negative feedback processes in blue) an initial perturbation ΔTs that results from radiative forcing of ΔF at the top of the atmosphere*

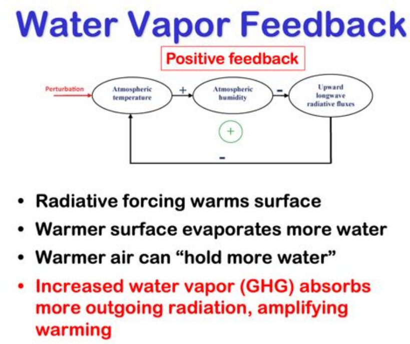
4.2.1 Positive Climate Feedback Processes

Positive feedback in the climate system means that temperature changes (either warming or cooling) are amplified. This happens when an initial temperature change causes changes in other variables that increase the radiation imbalance from the initial climate forcing.

You might be familiar with positive feedback used by some rock guitarists (Jimi Hendrix was the classic example). The signal from the guitar is amplified and blasted out of the monitor/speaker onstage. The guitarist holds the guitar up to the speaker so that the sound vibrations are quite literally ***fed back*** into the strings and pickups and that signal is also amplified, blasting back into the guitar. And so on. The effect can be electrifying!

Positive feedback always amplifies changes, making systems more variable.

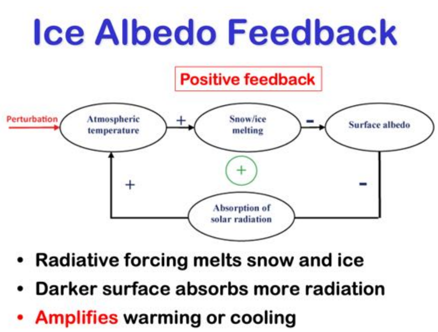
Water Vapor Feedback



When positive climate forcing (more radiation in than out) produces an initial warming of Earth's surface, evaporation from the oceans increases. Since water vapor is a greenhouse gas, the extra water vapor absorbs some of the outgoing longwave radiation (OLR), which results in even less radiation being emitted to space. This means that the response to an initial imbalance (warmer oceans with more evaporation) acts to increase the radiative forcing, which further increases the temperature. Of course, this also increase the evaporation and water vapor even more.

Conversely an initial cooling in response to negative climate forcing (more radiation out than in) will generally make more water vapor in the atmosphere condense to form clouds and precipitation. Removing water vapor from the atmosphere weakens the absorption of OLR, which tends to enhance the initial negative radiation balance. This causes further cooling which causes even more vapor to be removed, and so on.

Water vapor feedback is among the strongest and most important positive feedback processes int eh climate system. The average residence time of water vapor in Earth's atmosphere is only nine days, so water vapor feedback acts very fast to amplify warming or cooling of Earth's climate.



Ice/Albedo Feedback

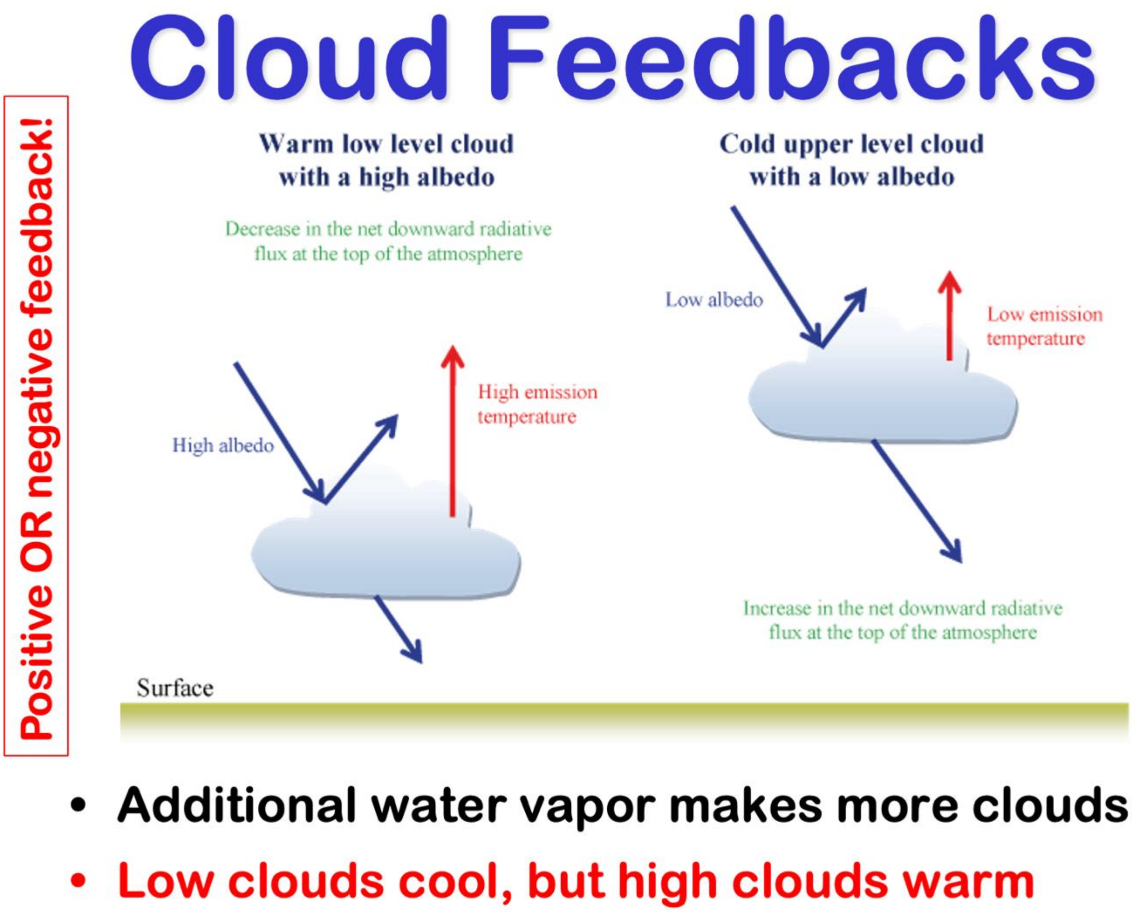
Initial warming of Earth's climate tends to melt snow and ice, reducing the planetary albedo and promoting additional absorption of incoming solar radiation. This additional absorbed energy increases temperature further, which melts more snow and ice, leading to more absorbed sunshine and so on.

Conversely an initial cooling tends to result in additional accumulation of snow and ice which increases planetary albedo. More sunlight is reflected to space, making the radiation balance even more negative (more radiation out than in). This further cools climate, producing more snow and ice and so on.

Ice/albedo feedback is a strong amplifier of climate change on Earth, but it acts more slowly than water vapor feedback. Seasonal snowpack develops and melts over periods of months. Sea ice, large glaciers, and continental ice sheets develop and melt over periods of decades to many centuries and have profound influences on global climate over these time scales.

High Cloud Feedback

Cloud feedbacks can be confusing, and indeed are an intense subject of professional climate research. Everybody knows that when a cloud obscures the Sun the surface temperature tends to drop, but we also know that cloudy nights tend to be warmer than clear nights. This dual nature of cloud forcing arises from the fact that clouds both cool climate by reflecting sunlight and warm climate by absorbing outgoing longwave radiation (OLR).



It turns out that whether the cooling (solar) effect or the warming (OLR) effect of clouds dominates is largely determined by the height of the cloud: low clouds tend to be more dense and optically thick (reflecting a lot of sunlight upward) whereas high clouds tend to be thin (letting sunlight in but blocking OLR).

This leads to a simple rule that's true most of the time:

"Low clouds cool the surface, but high clouds warm the surface."

4.2.2 Negative Climate Feedback Processes

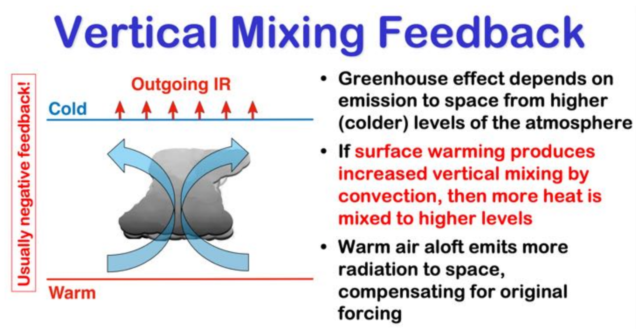
Negative feedback in the climate system means that temperature changes (either warming or cooling) are reduced. This happens when an initial temperature change causes changes in other variables that decrease the radiation imbalance from the initial climate forcing.

In everyday life the classic example of negative feedback is a thermostat. When the temperature of the room climbs above the thermometer set point, the furnace turns off to let the room cool down. Conversely when the temperature drops below the set point, the furnace turns on to warm the room back up.

Negative feedbacks always stabilize a system, holding it steady.

Vertical Mixing Feedback

We learned in Module 3 that one of the key factors that determine the strength of the Greenhouse Effect on climate is the drop in temperature with height. Selective transparency due to greenhouse gases like CO_2 and H_2O make the atmosphere opaque to outgoing longwave radiation emitted by the surface so the planet has to cool by emitting radiation from much higher aloft where temperatures are colder. The colder it is up at the level where OLR can escape to space, the less overall cooling the radiation can accomplish.



Vertical mixing by atmospheric convection carries heat form the warm surface to higher levels where emitted photons have a greater chance of escaping to space.

Near the surface, convective mixing is driven by buoyancy. Air in direct contact with the warm surface expands and becomes less dense than its surroundings. Consequently, the warm air rises and carries heat away from the surface. Cooler air falls to replace the air at the surface and the circulation continues. This process is greatly enhanced when water vapor in rising columns of air condenses to form clouds. Condensing water droplets in clouds release the latent heat that was required to evaporate the water in the first place, warming the air even more so it continues to rise even faster. Moist convection in convective clouds is one of the most efficient ways to get heat from the surface up to the level where it can be emitted to space as OLR.

Convective mixing generally increases in a warmer climate both because of the direct effect of additional heat on surface buoyancy and because of the increase in water vapor at warmer temperatures. ***Vertical mixing is therefore a strong negative feedback to the climate system*** because warming causes mixing to cool the surface and cooling makes the air more stable and less prone to mix. Less mixing carries less heat aloft where it can be emitted to space.

Low Cloud Feedback

In a warming climate with increasing water vapor, some of the additional water vapor can condense to form clouds. If the additional clouds are high in the sky and optically thin (like cirrus) they have relatively little effect on incoming solar radiation but trap a lot of IR. In that case we saw above that they act as positive feedback.

But low clouds tend to be much more dense and wet and optically thick. They block and reflect a lot of incoming solar radiation. At the same time, they emit thermal radiation upward from near the surface where it's relatively warm so they don't decrease the OLR by nearly as much as high cold clouds do.

Therefor low clouds act as negative feedback to the climate system. A warmer climate that produces additional low clouds will not warm as much as if those clouds hadn't formed.

4.3 Estimating Total Climate Sensitivity (Including all the Feedbacks)

Estimation of climate sensitivity to forcing (in degrees per W m^{-2}) is both much more difficult and much more useful when we include amplification and damping of climate change by positive and negative feedback processes.

Starting with the baseline (blackbody) sensitivity of 0.27 K / W m^{-2} , we expect this number to increase due to all the positive feedbacks and decrease due to all the negative feedbacks. As you might imagine, it's really hard to isolate just the effect of water vapor or high clouds or snow or mixing in the real world, so quantifying all these feedbacks is difficult.

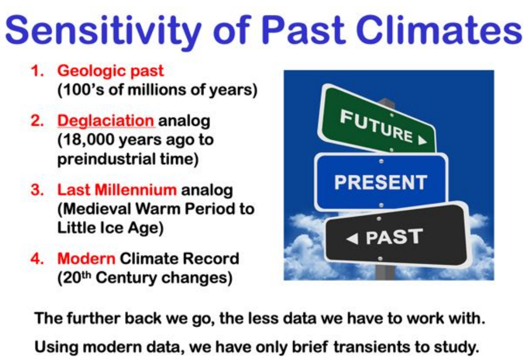
Climate scientists have two major classes of methods to analyze the total climate sensitivity:

1. By studying "paleoclimate analogs" (past climate change) and
2. By tracing all physical processes quantitatively using numerical models.

Sensitivity Estimates from Past Climate Change

Climate has changed tremendously in the distant past. We study past climate changes using evidence we find in rocks, soils, and ice. These "geologic proxies" record conditions in the distant past. Changes in temperature and moisture affect the chemical composition of minerals and other compounds deposited in sediments over many thousands of years. Over even longer periods of millions of years, plants and animals evolve and migrate to take advantage of different climate conditions in different places. Some of these ancient life forms are preserved as fossils in very old rocks, allowing geologists to reconstruct huge climate change over Earth history.

The idea is to try to reconstruct both the forcing (ΔF) and response (ΔT_S) so they can be divided to obtain an estimate of the total climate sensitivity $S = \Delta T_S / \Delta F$.



The big ***advantage of estimating total climate sensitivity from past climate changes is that all the feedbacks are included automatically***! Since it's the real Earth that's being used for inference, the real clouds, rainfall, albedo, and water vapor are guaranteed to be implicit in the measurements.

The big ***disadvantage of paleoclimate sensitivity reconstruction is that it's really difficult***! It's hard enough to reconstruct ancient temperatures using grains of pollen, fossil bones and teeth or the stable isotopes of mineral crystals. It's much harder to simultaneously suss out what the changes in solar radiation, OLR, and the net forcing must have been millions of years ago.

Usually, the further back in time we try to go with paleoclimate investigations, the less direct evidence we have available. On the other hand, most of the really huge changes in Earth's climate happened long ago, so this is attractive for estimation.

The data are really excellent over the past century or so. We have thousands of detailed records of temperature that go back to the mid-19th Century. Old measurements of radiation are less precise, but we can make century-scale estimates of climate sensitivity from the historical instrument record. The biggest problem with these data is that climate hasn't really changed all that much over the past few decades so small errors in either the numerator ΔT_S or the denominator ΔF can lead to large errors in climate sensitivity.

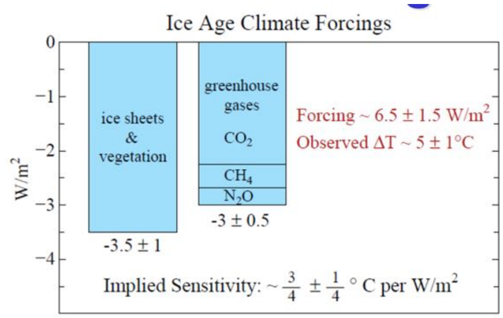
Since the Industrial Revolution, global mean surface temperature has warmed about ΔT_S = 1.2 Celsius. Increases in CO_2 and other greenhouse gases have produced radiative forcing of around 2.5 W m^{-2}, but this has been partly offset by reflective smog pollution. Climate scientists estimnate the net radiative forcing (greenhouse gases minus reflective smog) to be about ΔF = 1.5 W m^{-2}. So the modern climate data suggest a sensitivity of about $S = \Delta T_S/\Delta F = 1.2 \text{ K} / 1.5 \text{ W m}^{-2} = 0.8 \text{ K} / (\text{W m}^{-2})$.

Over the past few millennia, there have been some swings in climate associated with changing frequency of volcanic eruptions and slow changes in the brightness of the Sun. These changes are revealed in tree rings, snow and ice as well as in historical records of the "Little Ice Age (roughly the years 1400 through 1800) or the Medieval Warm Period (roughly the years 1000 through 1400). The trouble here is that the data are spatially sparse and imprecise.

From the height of the Medieval Warm Period around 1200 to the depths of the Little Ice Age around 1650, tree rings and other proxy data suggest a cooling of around ΔT_s = -0.8 Celsius (though there is some dispute as to whether this was truly global cooling or mostly too place in Europe). Sunspot counts, volcanic ash deposits, and other proxies suggest net radiative forcing of about ΔF = -1 W m^{-2} during those centuries. Again, this analysis yields an estimate of total climate sensitivity of $S = \Delta T_S/\Delta F = 0.8 \text{ K} / 1.0 \text{ W m}^{-2} = 0.8 \text{ K} / (\text{W m}^{-2})$.

Deeper in the past we have the amazing cycles of the Great Pleistocene Ice Ages (much more about this in Module 5 next week). These involve much stronger climate changes: around 5 Celsius of warming and cooling over periods of hundreds of centuries. The Last Glacial Maximum was "only" 18,000 years ago so there is actually quite a lot of physical debris left to be interpreted from the warming that followed this extreme cold snap. Ancient temperature changes are reconstructed using microscopic fossils in ocean bottom mud and other proxies. Changes in atmospheric CO_2 and other greenhouse gases can be measured directly in modern laboratories on microscopic bubbles in ice cores extracted from polar ice sheets.

Reconstructions of temperature across the globe suggest that over the period from about 18,000 years ago until 8,000 years ago, global mean temperatures warmed about $\Delta T_s = +5$ Celsius. Mapping of glacial deposits and sea ice show that the Earth's albedo decreased enough to add 3.5 W m^{-2} of absorbed solar radiation during deglaciation. Analysis of gas concentrations in the bubbles in ice cores show that CO_2 , methane, and other greenhouse gases increased to produce enhanced absorption of OLR in the amount of 3.0 W m^{-2}. Taken together, total radiative forcing during deglaciation was around 6.5 W m^{-2}, implying a total climate sensitivity of $S = \Delta T_S/\Delta F$ = 5 K / 6.5 W m^{-2} = 0.75 K / (W m^{-2}).



Source: Hansen and Sato (2011)

Over really deep time, there have been much larger changes in climate that are well documented in the geologic record. Dozens of events have been studies over hundreds of millions of years. Some of these dwarf even the Pleistocene Ice Age cycles, with temperature swings of 10 or even 20 Celsius. It is very difficult to get simultaneous records of CO_2 and other greenhouse forcing with timing precise enough to accurately assess the radiative forcing of most of these events.

Climate sensitivity derived from the historic record, the Little Ice Age, and deglaciation from the Last Glacial Maximum 18,000 years ago all show ***total climate sensitivity of around 0.8 K / (W m^{-2}). Note that this is around three times the baseline or blackbody climate sensitivity of 0.27 K / (W m^{-2}) derived form the Stefan-Boltzmann formula (equation 4-3). This implies that positive feedback dominates on timescales of decades to millennia.***

It's hard to imagine that the enormous climate changes documented in the geologic past could have occurred without strong positive feedbacks to amplify radiative forcing. By contrast, we know that over billions of years there has almost always been liquid water on Earth's surface, so climate has never "run away" to evaporate or freeze the entire oceans. This suggests that over even longer timescales negative feedbacks are also important to stabilize climate.

To put our modern global warming in perspective, it's remarkable to consider that ***deglaciation occurred due to radiative forcing of about 6.5 W m^{-2}. That's less than the upper bound of radiative forcing projected by 2100 under moderately high emission scenarios. By 2300, modern radiative forcing could be twice what caused deglaciation***.

Following the Last Glacial Maximum 18,000 years ago global temperature warmed about 5 Celsius over 10,000 years. That's a warming rate of 0.05 ºC per century, and it completely changed the geography of the world! The Bering Land Bridge was submerged, sundering the New World from the Old. Britain was cut off from mainland Europe. The flow of water through the Straits of Gibraltar reversed direction. By contrast, we could easily see that much warming in a single century, which is ***100 times the rate of deglaciation***!

Sensitivity Estimates Using Physics & Models

An alternative to estimating climate sensitivity from past climate change is to use physics and mathematics to try to calculate all the climate feedback processes by brute force. This is the job of climate models, which we will consider in much greater detail in Module 7.

The idea with climate models is to consider local changes in climate variables (temperature, water vapor, wind, pressure, rainfall, clouds, radiation, etc) on a three-dimensional grid using known laws of physics, then march the solution through time every few minutes for many centuries. Climate models produce maps of each variable over time. Maps of the past can be quantitatively compared against observations to test and improve the fidelity of simulated climate. Projections of the future can be tested under varying assumptions for future population, economic development, greenhouse gas emissions, and so forth.

A key advantage of using physics and calculating the feedbacks with numerical models is that we can gain mechanistic understanding of the processes that control climate change. The key disadvantage is that unlike analyzing Earth's real climate in the past, our models can be wrong!

Remarkably, estimates of total climate sensitivity (including all the feedbacks) using modern climate models agree very well with estimates of climate sensitivity derived from past climate change. Total sensitivity is estimated to be around 0.8 K / (W m^{-2}) by both sets of methods.

Before we consider comprehensive comparisons of climate sensitivity estimates across many methods, let's look at the specific case of radiative forcing of climate by CO_2 and other greenhouse gases.

4.4 Climate Forcing by CO_2 and other Greenhouse Gases

Swedish chemist Svante Arrhenius famously wrote in 1896 that

> *"if the quantity of carbonic acid increases in geometric progression, the augmentation of the temperature will increase nearly in arithmetic progression."*

By this, Arrhenius meant that the ***effect of CO_2 on radiation and temperature follows a "law of diminshing returns."***

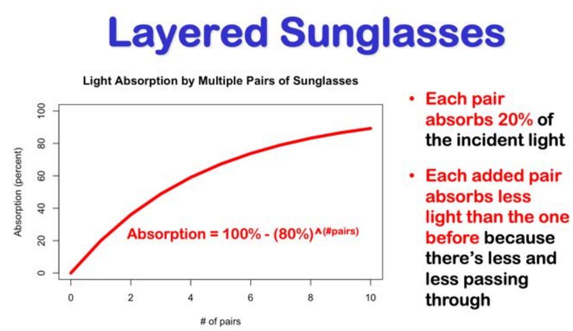
In plain language, each 100 ppm of additional CO_2 warms the climate less than the addition of 100 ppm that precedes it. This may sound a bit weird but if you think it through it is just common sense, and in fact ***this is the way all absorption works (not just for CO_2).***

4.4.1 Ridiculous Eyewear Fashions

Imagine that you have a pair of sunglasses that absorbs 20% of the incoming light. You put them on and your view of the world is 80% as bright as it was without the glasses.

Now imagine you put on a second pair of identical sunglasses (yes, this would look stupid but bear with me here). The second pair also absorbs 20% of the light that hits them, but since the light reaching the inner pair is only 80% of the outdoor light, the absolute amount absorbed by the inner pair is less than the amount absorbed by the outer pair. If the ambient light outdoors is 100 units of brightness, then the outer pair of sunglasses absorbs 20 units and lets 80 units through but the second (inner) pair absorbs only 20% of the 80 units (16 units) that got through the outer pair.

If you add a third pair of 20%-absorbent shades, it will absorb less light than the second pair because there's less light left over for them to absorb. The third (innermost) pair will now absorb 20% of 100 units x (100%-20%) x (100%-20%) = 13.8 units of brightness.



You can take this concept to the level of absurdity. Each additional pair of sunglasses makes the scene darker by the same relative amount, but because there's less light left for them to absorb each additional pair affects the final brightness reachng your eyes less than the pair before them.

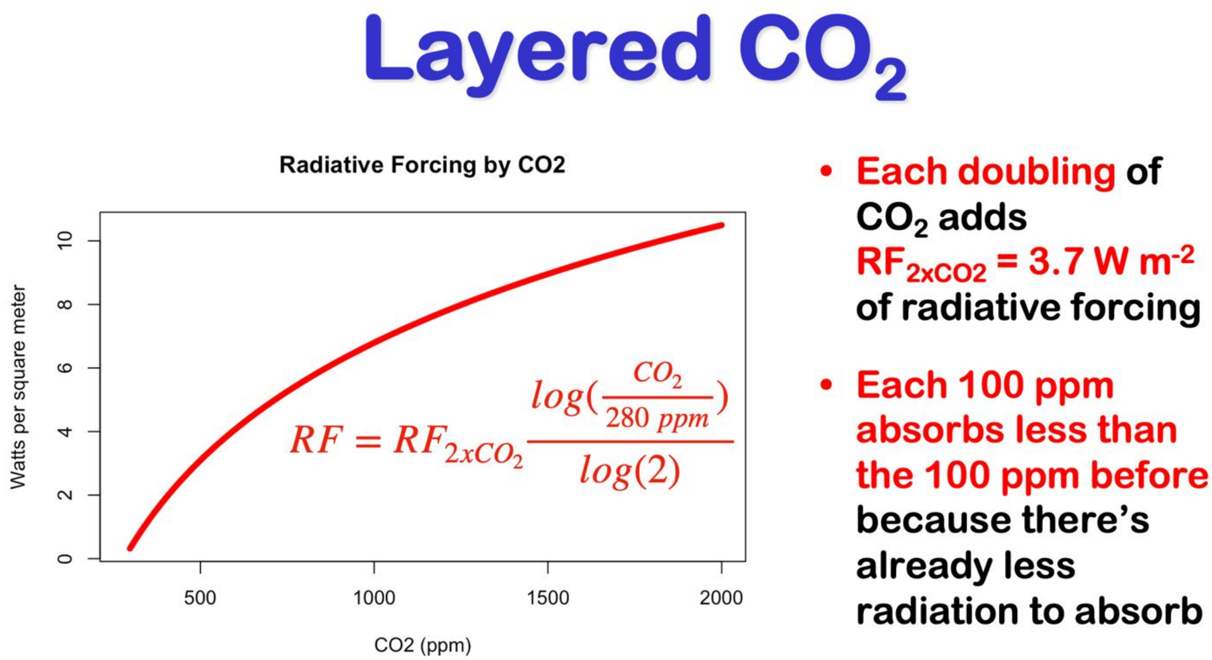
Mathematically, the amount of light absorbed by the nth pair of sunglasses is calculated by a power law: we raise the transmission of the lenses to the power of the number of pairs. This produces the classic flattening curve above, known in plain language as "diminishing returns" or mathematically as a ***logarithmic curve***. Each pair of layered sunglasses absorbs less light, but there is no final number beyond which adding another pair won't absorb 20% of what's left.

4.4.2 CO_2 Acts Just Like Sunglasses for OLR

Greenhouse absorbers act on Earth's outgoing longwave radiation (OLR) in exactly the same way sunglasses act on bright daylight. Each addition of the same amount of greenhouse gas (GHG) absorbs less than the amount added previously, so progressive additions absorb less and less radiation because there's less available after passing through what was there before.

Just as we found for layered sunglasses (and for precisely the same reason), the absorption of OLR by CO_2 follows a logarithmic curve which tends toward but never reaches total saturation.

It's customary to refer to the radiative forcing of climate by CO_2 in terms of how many doublings of CO_2 concentration in the air. Each doubling of CO_2 absorbs 3.7 W m^{-2} of OLR (holding temperature constant). We write $RF_{2xCO2} = 3.7$ W m^{-2} to indicate this. This is precisely what Arrhenius meant 125 years ago when he wrote that geometric additions of CO_2 add arithmetically to warming.



Quantitatively, we can easily calculate the radiative forcing of climate relative to preindustrial conditions due to any level of CO_2 in the air:

$$RF(CO_2) = RF_{2xCO2}\frac{log\ (\frac{CO_2}{280ppm})}{log\ (2)} \tag{4-4}$$

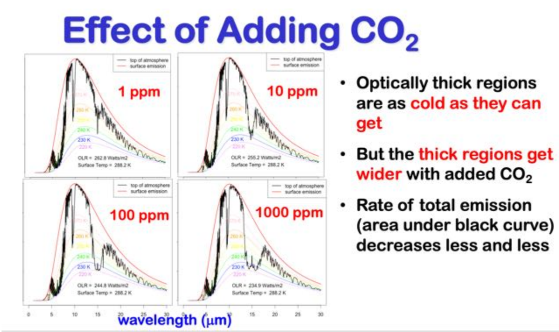
In equation (4-4) $RF_{2xCO2} = 3.7$ W m^{-2} is the radiative forcing per doubling of CO_2 and 280 ppm was the concentration of CO_2 in the atmosphere in preindustrial times (prior to about the year 1800). The *log*(2) in the denominator indicates that we're counting doublings. It doesn't matter whether you use log_{10} or natural log (*ln*, log_e) for equation (4-4) as long as you use the same base in both the numerator and denominator.

Some of you are old enough to remember the little incandescent light bulbs we used to use as nightlights to help kids find the bathroom at night. These are still available but quickly being replaced by much more efficient LEDs. They were 4 watt bulbs, so they emit almost exactly the same amount of heat as is absorbed by every square meter of the world when CO_2 is doubled (3.7 W, holding temperature constant).

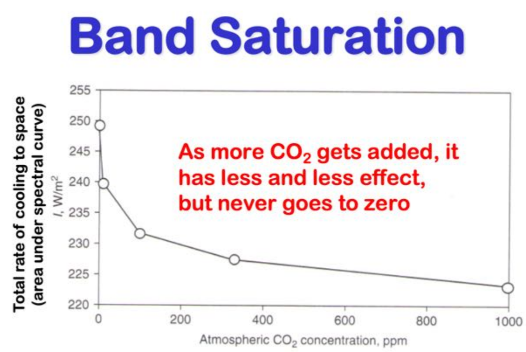
We can think of ***every doubling CO_2 as adding one of those old-fashioned night light bulbs to every square meter of the world***. Before the industrial revolution, there weas about 280 ppm of CO_2 in the air. At 560 ppm (2x preindustrial CO_2), the world will warm up as much as if we turned on a night-light in every square meter of the world. At 1020 ppm (4x preindustrial CO_2) we will have added 2 night-lights to every square meter. At 2040 ppm (8x preindustrial CO_2) we will have added 4 night-lights to every square meter. And so on.

There are two pieces of bad news here:

1) There's no way to turn the night lights back off except by *removing* CO_2 from the air; and
2) There's no point at which adding CO_2 to the air doesn't add more night-lights.



As we saw in Module 3, adding more and more CO_2 to the air lifts the effective altitude from which OLR photons escape to space higher and higher. Beyond a certain altitude in the stratosphere (about 10 or 12 km) the temperature no longer decreases with height so absorption at the central wavelength can't decrease any more. But instead adding more CO_2 widens the bite of wavelengths of OLR that gets cut out by each CO_2 absorption band. It takes more and more CO_2 to absorb the same amount of OLR (band saturation), but the amount absorbed never goes to zero no matter how much CO_2 is added.



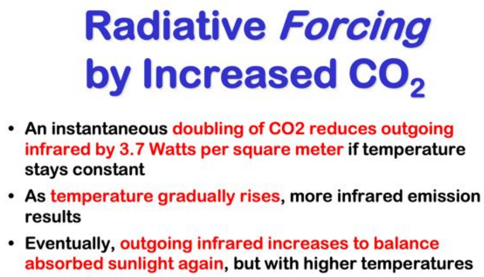
4.4.3 Equilibrium Warming by CO_2

Now we can combine the concept of climate sensitivity (degrees per W m^{-2}) with the logarithmic radiative forcing from CO_2 and other GHGs. We know from both past climate change and by brute-force calculation of climate feedback processes through numerical models that total climate sensitivity is about 0.8 K / (W m^{-2}), and we have seen that each doubling of CO_2 adds 3.7 W m^{-2}.

Therefore, assuming that Earth's climate is just as sensitive to Watts absorbed by CO_2 as it is to all other forms of heat, surface temperature should increase

$$S_{2xCO2} = (3.7 \text{ W m}^{-2} \text{ per } 2xCO_2) \times (0.8 \text{ Kelvin per W m}^{-2}) = 3 \text{ Kelvin per } 2xCO_2 .$$

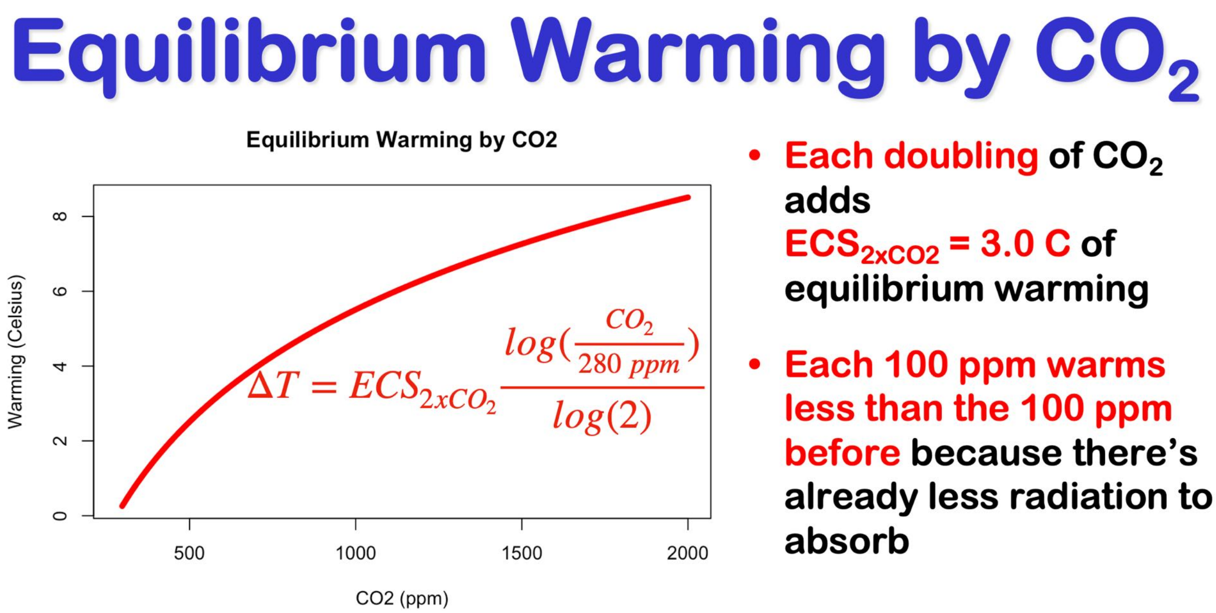
This is the same climate sensitivity we derived in section 4.3, but expressed in a different unit (degrees per doubling of CO_2 rather than degrees per W m^{-2} of radiation). It's probably more common to see climate sensitivity written in terms of CO_2 than W m^{-2}, particularly with respect to modern climate change. But it's useful to remember that climate is equally sensitive to any change in the radiation balance of the planet. All the feedbacks (water vapor, clouds, snow and ice, vertical mixing, etc) act on the radiative forcing from CO_2 just as they would act on a change in the brightness of the Sun or a cloud of volcanic dust or any other forcing.



It's customary to write that the ***Equilibrium Climate Sensitivity*** (ECS) due to CO_2 is

ECS = 3 Kelvin per 2xCO_2 = 3 °C per doubling of CO_2 .

The important word "equilibrium" in ECS refers to the idea that it takes time (several decades at least) for climate to equilibrate with radiative forcing. In other words, if CO_2 were to double instantaneously the climate would warm over a period of decades, and eventually the OLR would increase with temperature enough to balance the extra 3.7 W m^{-2}. If ECS = 3 °C per doubling of CO_2, then the new surface temperature would be 3 °C warmer after these decades had passed than it was before CO_2 doubled.



A related quantity is the ***Transient Climate Response (TCR), which indicates how much warming will occur at the time CO_2 reaches double its preindustrial value*** (560 ppm compared to 280 ppm in 1800). The TCR depends on how quickly CO_2 rises and other details, but it is certainly less than ECS because the climate won't have time to adjust to rising CO_2. Most studies estimate ***TCR is about 2 °C*** at the time CO_2 hits 560 ppm (compared to 3 °C per doubling for ECS).

4.5 Comparing Climate Forcing, Response, & Sensitivity

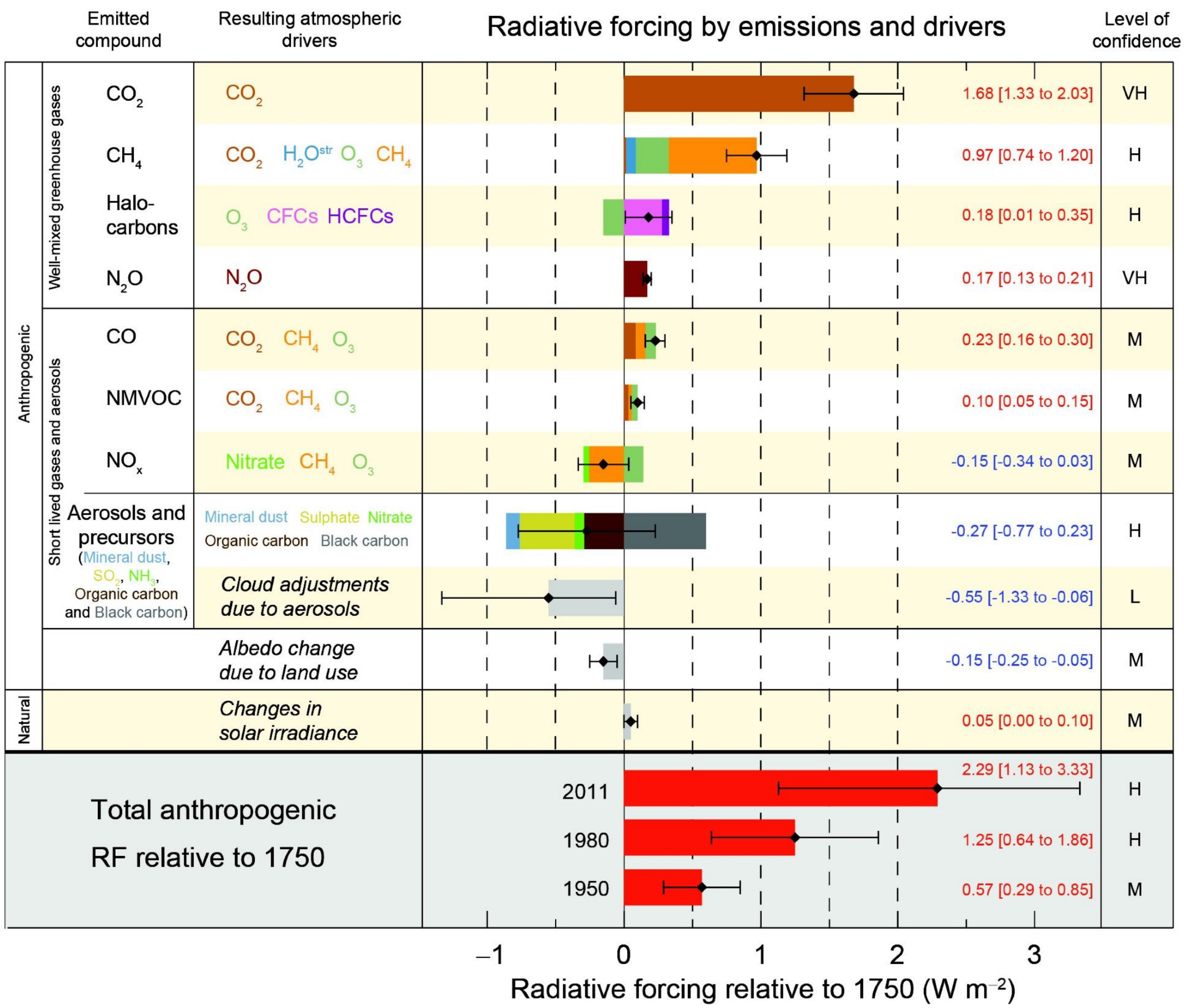


Figure 4-4 Comparison of radiative forcing for many atmospheric constituents. IPCC 6th Assessment 2022

Radiative forcing of modern climate has been calculated for many gases and other processes relative to preindustrial conditions (Fig 4-4 above). By far the biggest contribution is from CO_2 (about 1.5 W m^{-2}) and methane (CH_4, about 1 W m^{-2}). These GHG forcings are offset by a large and much more uncertain negative forcing by reflective particles derived form air pollution (smog, labeled "aerosols and precursors in the figure). The effects of air pollution are especially hard to nail down because the direct impact on the albedo of air (-0.4 +/- 0.5 W m^{-2}) is amplified by an even bigger impact on the albedo of clouds (-0.5 W m^{-2}). It's this ***uncertain offset by reflective smog that makes total anthropogenic (human-caused) climate forcing so uncertain (2 +/- 1 W m^{-2}).*** By comparison, the effects of human land use change (primarily deforestation, about -0.1 W m^{-2}) and changes in the brightness of the Sun (about +0.05 W m^{-2}) are much smaller than either the warming effects of GHGs or the cooling effects of reflective smog.

Comparing estimates of total equilibrium climate sensitivity (ECS) across literally hundreds of studies using dozens of different methods (paleoclimate, theoretical, observational) and find that there is broad agreement on a sensitivity of about 3 °C per doubling of CO_2. The uncertainty in this critical number is stubboirnly high and hasn't improved in decades.

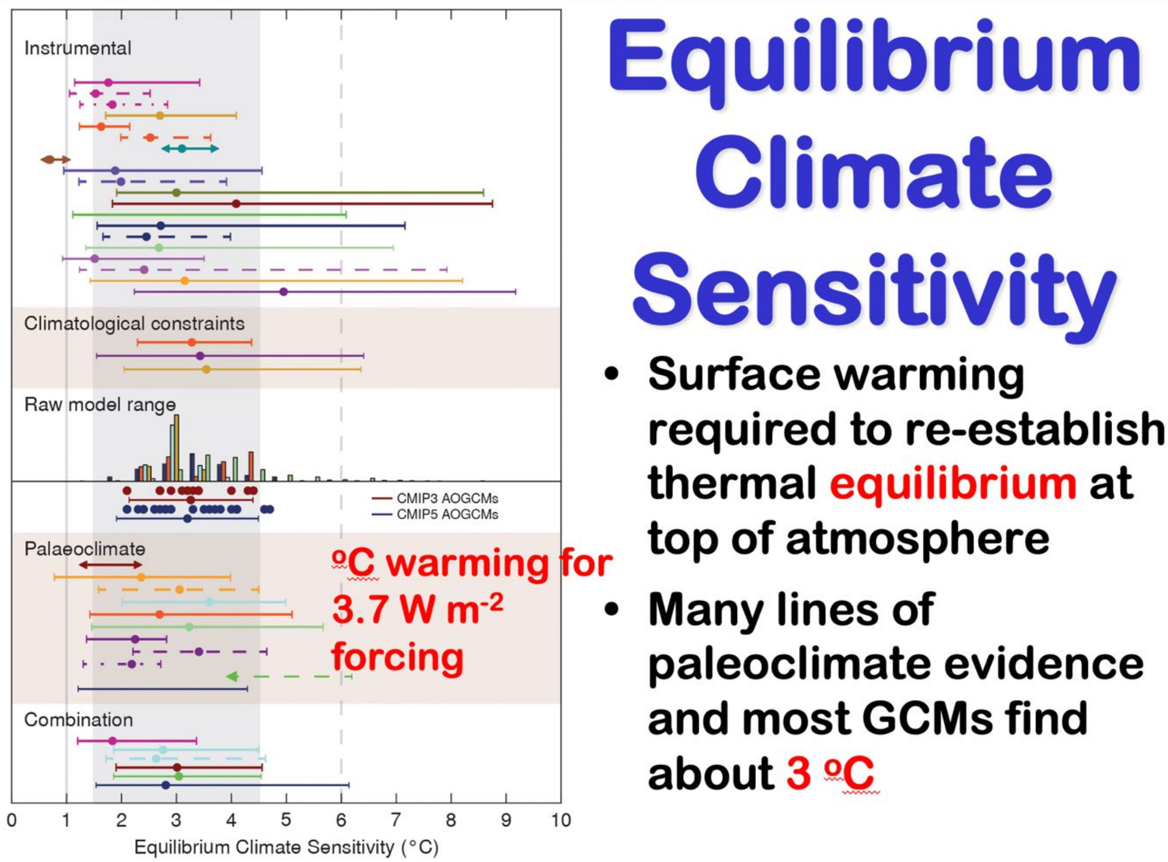


Figure 4-5: Comparison of ECS estimates across many studies and methods. IPCC AR5 (2013)

Similar comparisons have been made for the smaller Transient Climate Response (TCR), which is the amount of warming expected at the time CO_2 reaches 560 ppm, some decades before climate reaches equilibrium with radiative forcing. This number is generally estimated at between 1.5 °C and 2.5 °C (see figure below).

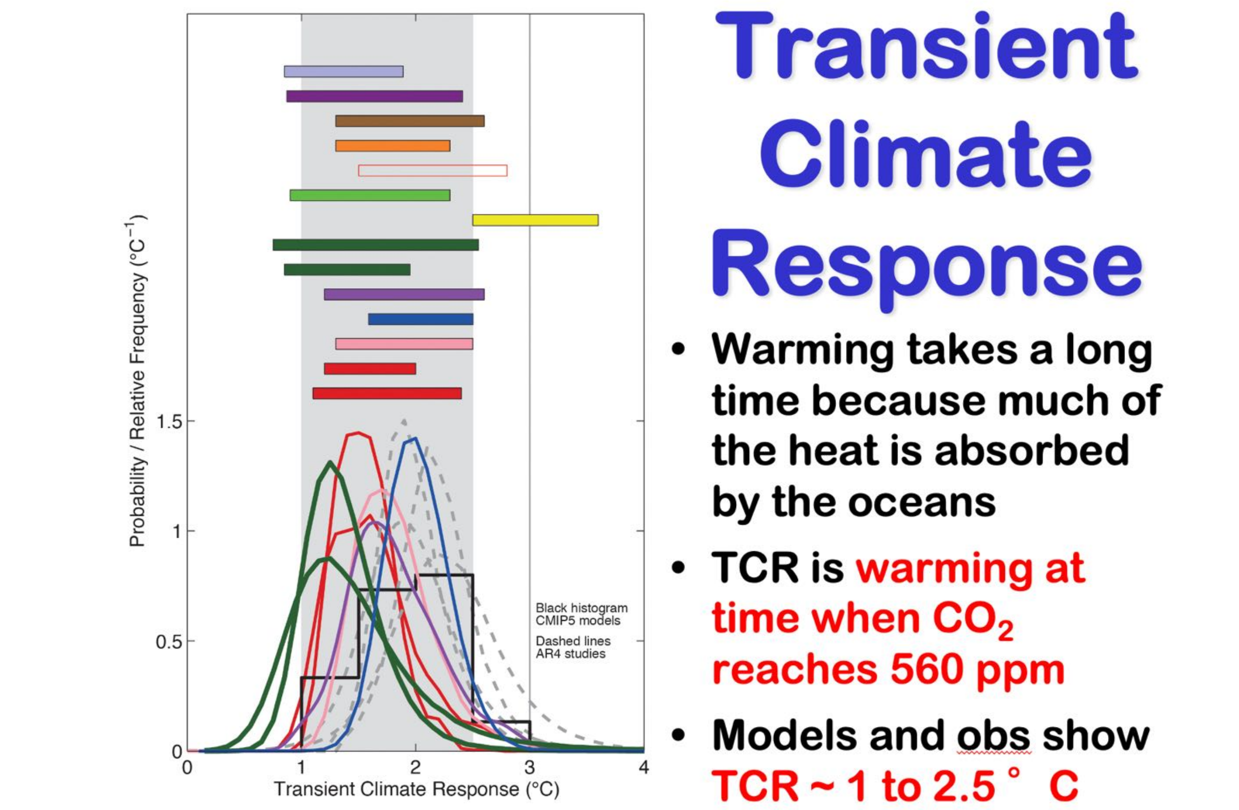


Figure 4-6: Comparison of TCR across many estimates (IPCC AR5, 2013)