

An Analysis of Radiative Equilibrium, Forcings, and Feedbacks

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Abstract

Changes in the Global Mean Surface Temperature of the planet are determined by radiative forcings and feedbacks. Anthropogenic activities since the industrial revolution, such as the combustion of fossil fuels, have changed the chemistry of the atmosphere which creates a planetary energy imbalance that favors a global warming situation. The extent to which the climate will change in the future depends extensively on radiative feedbacks which either amplify or dampen the imposed perturbation on the climate system. I summarize the current science of forcings and feedbacks, along with basic radiative balance concepts. It is shown that the inherent nature of feedbacks makes a very high sensitivity difficult to rule out, but climate sensitivity is likely between 2 and 4.5 °C per doubling of CO₂. Further background is given on the climate change issue and current and projected trends.

1) Introduction

In radiative equilibrium, the amount of solar (shortwave) radiation absorbed by the planet equals the outgoing infrared (longwave, or thermal) radiation that the planet emits back to space (OLR) at the top-of-atmosphere. Because the OLR is a function of temperature, when the radiation balance of the Earth is perturbed (e.g., by increased solar irradiance, changes in greenhouse gases, etc), the global surface temperature will change as the planet adjusts to a new radiative equilibrium state. This section describes briefly how Earth's no-atmosphere temperature is set by radiative balance.

The absorbed sunlight on the planetary sphere declines with latitude as you move poleward, and varies with time as the Earth moves in orbit. The difference between the distance of the sun to the equator, and the distance from the sun to the poles is negligible, so the intensity difference is in the angle of incoming sunlight. The incoming solar irradiance s at some specific area as a function of location \mathbf{x} and time \mathbf{t} that originates from the solar constant \mathbf{Q} (in Watts that pass through each square meter of the planetary sphere) is,

$$s(x, t) = \cos(\theta(x, t))Q(t) \quad 1$$

Multiplying by the full area factor and integrating over a planetary sphere that is lit on only one side, we find the total incoming energy,

$$E_{in} = \int_0^{\frac{\pi}{2}} Q(t) 2\pi r^2 \cos \theta \sin \theta d\theta = Q(t) \pi r^2 \quad 2$$

Of the incoming solar radiation, a certain ratio α is reflected back out to space by “bright objects” (e.g., clouds, ice and snow, etc) without thermodynamic interaction with the climate system. This property is the so-called planetary *albedo*. All objects also emit their own radiation (planetary bodies do so in the infrared spectrum) in accord with the Stefan-Boltzmann law, $E_{out} = \epsilon \sigma T^4$ where ϵ is the emissivity (the ratio of how well an object absorbs/emits to a perfect blackbody), σ is a constant ($5.67 \times 10^{-8} \text{ W/m}^2 \text{ K}^4$) and T is the absolute temperature.

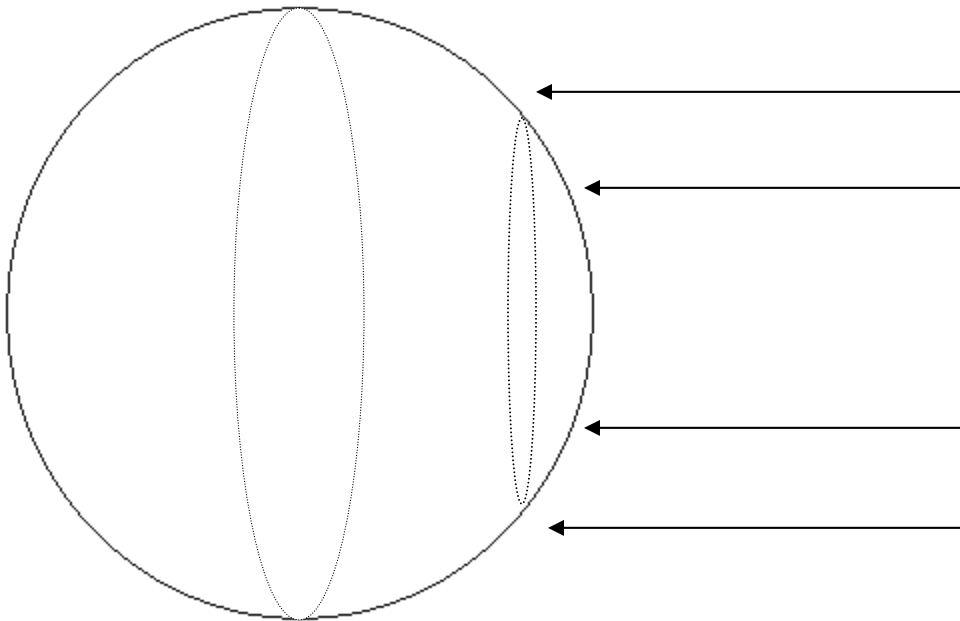


Fig 1- Exposed Earth face. When sunlight strikes the Earth it comes from one direction and the planet makes a circular shadow.

Although the planet takes in an influx of energy equal to the product of the sunlight intensity and the area of a circle, it radiates heat away to space like a full sphere since energy leaves the planet in all directions. Internal heating on the terrestrial planets is negligible. It follows from above that in radiative balance,

$$Q(t)(1-\alpha)\pi r^2 = 4\pi r^2 \epsilon \sigma T^4 \quad 3$$

Solving for the effective temperature, where the Earth is a blackbody with no atmosphere ($\epsilon = 1$),

$$T_{eff} = \sqrt[4]{\frac{Q(t)(1-\alpha)}{4\sigma}} \quad 4$$

Hereafter, allow

$$\frac{Q(1-\alpha)}{4} \equiv Q_{in} \quad 5$$

The current solar constant is approximately 1,370 Watts per square meter (W m^{-2}), and the planetary albedo is 0.3 (i.e. 30% of incoming sunlight is reflected away), so the day lit side of the planet currently absorbs roughly 240 W m^{-2} . This makes $T_{\text{eff}} \sim 255 \text{ K}$ ($0 \text{ }^\circ\text{F}$) (Trenberth et al., 2009; in press). Without an alternative heat source for the planet or an infrared absorbing atmosphere, the average planetary temperature will tend toward and not exceed the effective temperature. If we do indeed allow Q_{in} and σT^4 to vary in time and remain global averages, then,

$$c \frac{dT}{dt} = Q_{in} - \sigma T^4 \quad 6$$

Where c is the specific heat capacity of the ocean, or $1.7 \times 10^{10} \text{ J m}^{-2} \text{ K}^{-1}$.

Planetary bodies are surrounded by a vacuum of space, and thus do not gain or lose heat by non-radiative processes. However, conduction and convection are important terms for the *surface* energy balance and regulate the surface-atmosphere temperature gradient and establishment of weather systems. The distribution of heat around the planet by atmospheric and ocean circulation helps moderate substantial temperature gradients both vertically and between the poles and equator. However, globally-induced climate changes through radiative forcings are defined at the top-of-atmosphere and can only occur with changes in Q_{in} (either by Q or α) or by trapping outgoing infrared heat and affecting the OLR, with the surface energy budget dragged along.

All of this is driven by the incoming sunlight. If Q_{in} fell to zero then the planetary temperature would eventually drop to nearly absolute zero (more precisely, the microwave background of about 2 K). However it is obvious from the result that $T_{\text{eff}} = 255 \text{ K}$ that Q_{in} is not the only relevant factor in global temperature. If this were the case, the oceans would be sufficiently frozen over.

It was assumed that an escaping infrared photon to space can exit freely and without trouble, and so the temperature where radiation is lost to space is that of the surface. In reality, the atmosphere serves as a good absorber of outgoing energy. Observations indicate that the global average temperature is roughly 288 K, and the atmosphere being partially opaque to outgoing infrared radiation is the reason for the 33 K ($33 \text{ }^\circ\text{C}$, $59 \text{ }^\circ\text{F}$) difference.

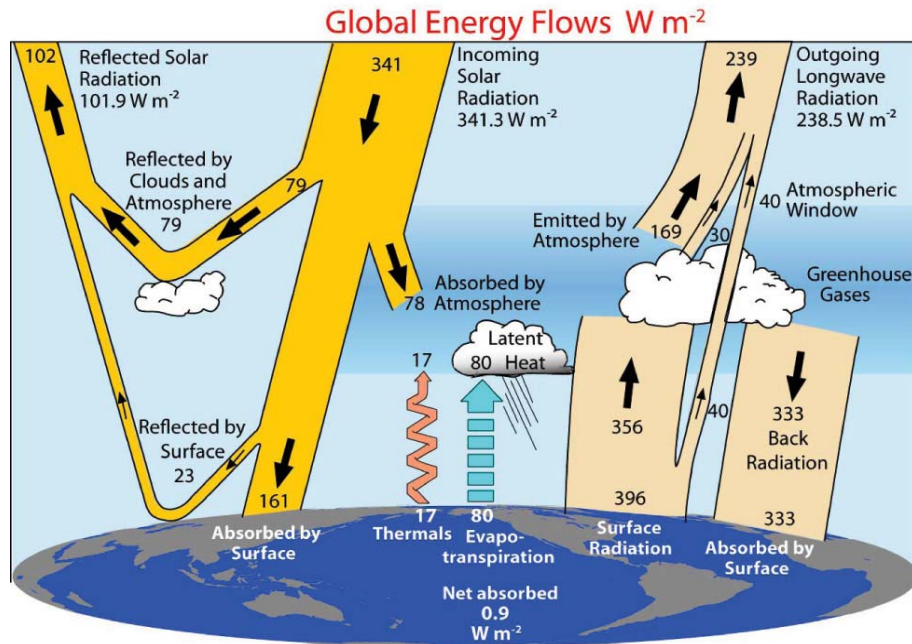


Fig 2- Earth's Global Energy Balance illustration. Re-Printed from Trenberth, Fasullo, and J. Kiehl, 2009

2) Atmospheric Greenhouse Effect

The fact that the existence of an atmosphere can inhibit the radiation loss to space and provide further warming to the surface originated almost two centuries ago (Fourier 1827) and the effects of changing the greenhouse concentrations of carbon dioxide (CO_2) was first quantified almost 70 years later (Arrhenius 1896).

The gas density of a planet with an atmosphere declines with height as gravity pulls molecules to the surface, and with less weight overhead as you go up, there is a resulting expansion and cooling of air molecules with altitude. The temperature decline with height is often expressed as the adiabatic lapse rate, which is set by convection and other vertical mixing processes. The adiabat is close to $-10\text{ }^\circ\text{C}/\text{km}$ in the polar and drier areas of the Earth. In moist areas, however, some cooling is offset by the latent heat release of condensing water vapor and the adiabat can reach $-4\text{ }^\circ\text{C}/\text{km}$ in the tropics.

If we now introduce some gas in the lower atmosphere that is transparent to incoming solar radiation, but opaque to outgoing terrestrial radiation, that gas will absorb and emit the infrared photon in a random direction, or it may transfer its energy to a neighboring molecule and warm the air where that gas resides.

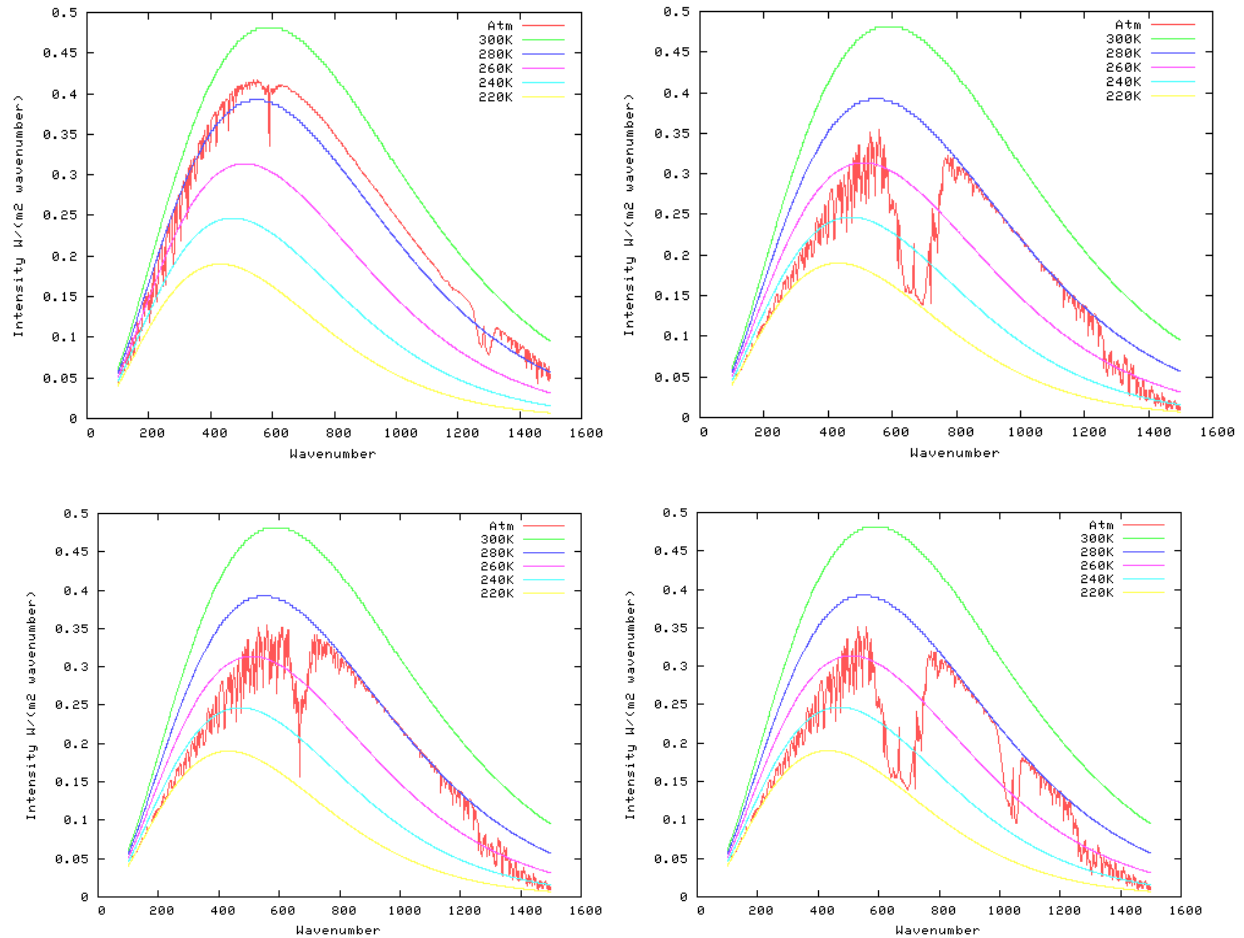


Fig 3- Pictorial depiction of the OLR being cut off at different frequencies. a) no atmosphere b) Just 2 ppmv of CO2 c) 2x modern CO2 concentrations d) Modern concentrations of CO2, methane, ozone. Plots created using online MODTRAN model

Radiation to space can come from virtually all parts of the atmosphere, but the bulk of it comes from the $\tau=1$ level roughly 5 km above the surface. Most of the radiation emitted from below $\tau=1$ is absorbed before it can get out to space, and radiation emitted from the uppermost levels of the atmosphere is small because the absorption and emissivity there is small, and little flux is emitted. A thorough treatment of this topic is in R.T. Pierrehumbert, Principles of Planetary Climate.

The effect of this absorbing layer is to reduce the OLR for a given temperature, so the planet heats up until radiative equilibrium is re-established. This occurs when $Q_{in} = OLR$, but this outgoing term is reached higher in the cooler, thinner parts of the atmosphere and not at the surface. With further addition of a greenhouse gases, the $\tau=1$ level is shifted to higher altitudes where it is colder, and hence by Stefan-Boltzmann, emits much more feebly. The temporary result is for $Q_{in} > OLR$ until a new equilibrium is established at a higher temperature.

CO₂ absorbs infrared radiation and renders the atmosphere completely opaque between 14 and 16 μm , and partly opaque some distance on each side of those wavelengths. Increased CO₂ acts to further widen this “bite in the spectrum,” and narrow the “atmospheric window” where radiation escapes easily and is now emerging from higher, colder layers (Fig 3).

The greenhouse warming is thus (Hansen et al 1981),

$$T_{surface} = T_{eff} + \Gamma H_{\tau=1} \quad 7$$

Where Γ is the lapse rate and H is height. Because the atmosphere radiates both up and down, the atmospheric energy budget is,

$$2\varepsilon\sigma T_{atmosphere}^4 = \varepsilon\sigma T_{surface}^4 \quad 8$$

Because the surface is heated by both incoming sunlight and downwelling infrared radiation from the atmosphere,

$$\varepsilon\sigma T_{surface}^4 = Q_{in} + \varepsilon\sigma T_{atmosphere}^4 \quad 9$$

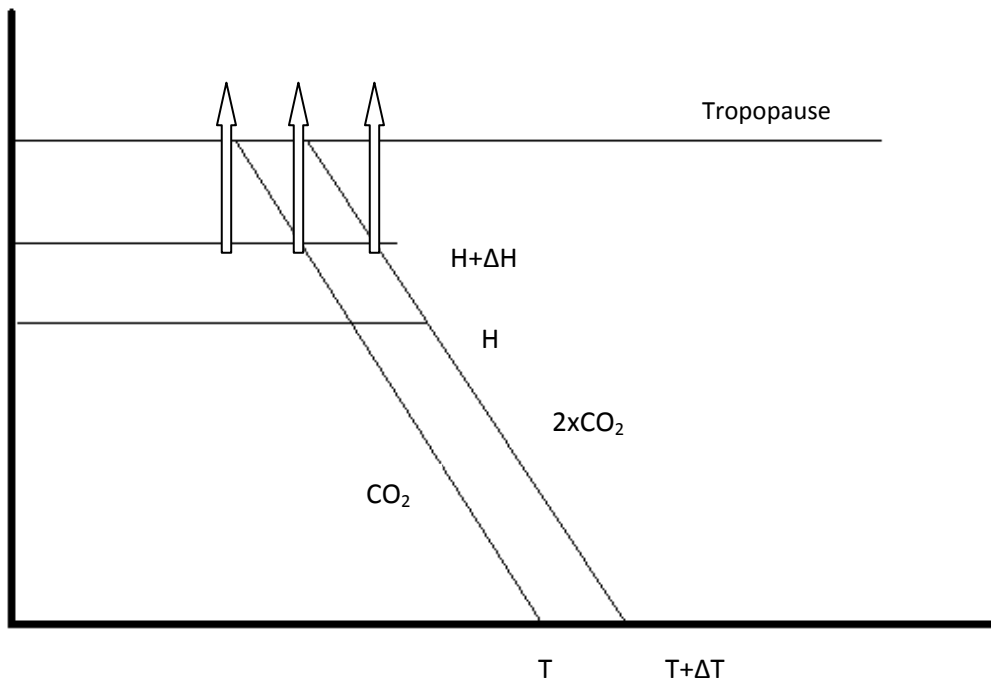


Fig 4- Illustration of heightened effective emission level (H, or $\tau=1$) with a doubling of CO₂. Concept from Held and Soden 2000.

Harries et al., 2001 have provided experimental proof for the enhanced greenhouse effect from orbit by comparing the outgoing long-wave radiation spectra from the Earth in 1970 and 1997, which shows an imprint of primarily CO₂ and CH₄ forcing, along with other anthropogenic-induced compounds.

3) Radiative Forcing

Factors that affect climate change are usefully separated into radiative forcings (RF) and feedbacks (Ramaswamy et al., 2001). An RF is an energy imbalance imposed on the climate system, while a climate feedback is an internal response to the climate change which amplifies or dampens the climate response.

The RF concept has received criticism (discussion in NAS 2005) for the limited ability to account for climate changes which are large at the local or regional level but have little global impact, changes in the hydrologic cycle, and many aspects of aerosols (e.g., they can have a positive top-of-atmosphere RF but still cool the surface). Nevertheless, RF's are a useful first-order way to quantify the impact of a given change to boundary conditions in the climate.

By definition, climate is a 30-year period of time, so to speak of climate change the forcing must typically be persistent over decades. If the forcing is persistent and large enough, it may produce a temperature trend that can be detectable against natural, internal variability. Volcanic eruptions cause a significant cooling effect by emitting volcanic ash and sulfur dioxide, which oxidizes into sulphate aerosols and reflect incoming sunlight back to space. However, the effects of this are short-lived (a year or two) so typical individual eruptions would not induce a climate change, but a long-term trend in volcanic activity could.

Human activities such as fossil fuels emissions and deforestation since the industrial revolution have increased the greenhouse concentrations of the atmosphere. The anthropogenic-induced warming trend over the last century has emerged from the background noise of natural variability (e.g., Meehl et al., 2004; Allen et al., 2006; Hegerl et al., 2006). CO₂ and methane (CH₄) changes from anthropogenic activities have been the two largest RF's since pre-industrial times (IPCC 2007, see Figure 6 of this document). CO₂ concentrations in the atmosphere are currently at 385 parts per million by volume (ppmv) and have undergone a 37% increase since pre-industrial times from 280 ppmv (Keeling et al., 2008). Ice core records confirm that CO₂ concentrations are higher, and rising at a much faster rate than anytime in at least 800,000 years (Lüthi et al., 2008, see Figure 5 of this document), and likely back several million years.

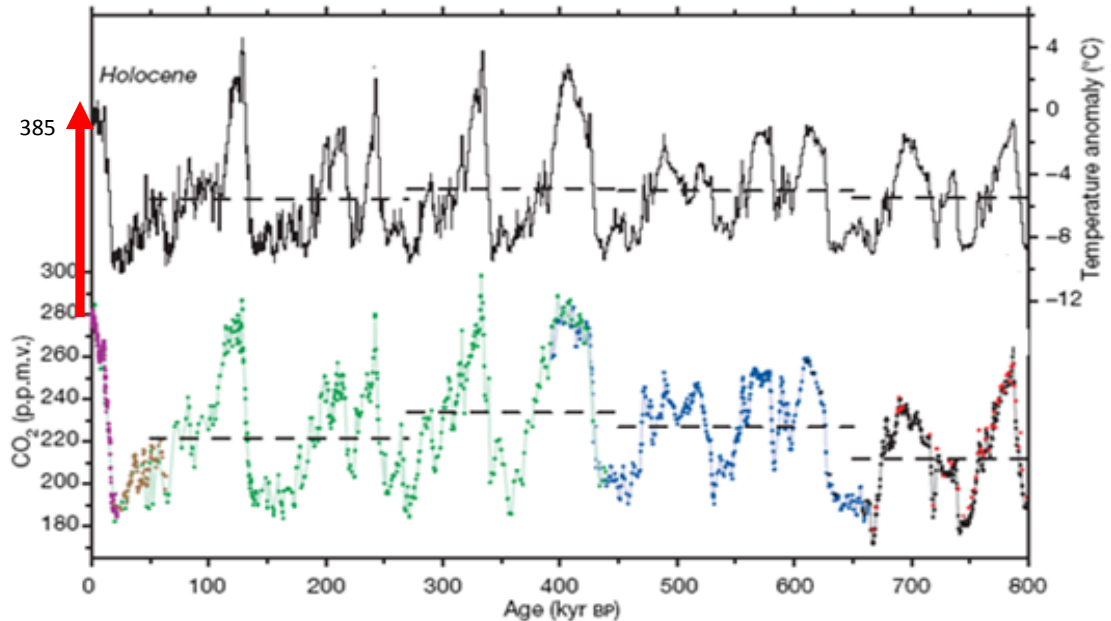


Fig 5- Compilation of CO₂ records and EPICA Dome C temperature anomaly (relative to mean over last millennia) over the past 800 kyr. Red line denotes industrial rise of CO₂. Modified from Figure 2 of Lüthi et al 2008.

CO₂ will receive key attention here since it represents the largest single RF since pre-industrial time, and its concentrations are expected to rise faster and provide the most warming potential over the coming century. The RF for a given change in CO₂ (Myhre et al., 1998) is,

$$F = 5.35 \ln\left(\frac{C}{C_0}\right) \quad 10$$

Where C and C₀ are the final and initial concentrations of the gas in ppmv. Because of the logarithmic nature, the response to successive increments of CO₂ changes become progressively smaller, and each doubling produces the same forcing (e.g., going from 50 to 100 ppmv has the same effect as going from 1000 to 2000 ppmv). This is a fortunate constraint since small changes in carbon dioxide will not induce radical shifts in climate that could frequently make Earth inhabitable; however, the changes are still large enough for relevant socio-economic concern. For a doubling of CO₂, the RF is approximately 3.7 W m⁻². This is similar to a 2% increase in total solar irradiance.

The forcing strength of individual gases depends upon where on the electromagnetic spectrum it absorbs radiation, its overlap with other gases that absorb in similar wavelengths, and current amounts in the atmosphere. The logarithmic effect no longer holds at extremely low or extremely high concentrations. Changes in CH₄, for instance, induce an RF greater than CO₂ by a factor of almost 30 at current Earth-like conditions, as incremental changes are not currently on a logarithmic scale. If CO₂ existed at the low concentrations of CH₄ and vice versa, the CH₄ forcing would be very small and CO₂ would produce a much larger effect.

RADIATIVE FORCING COMPONENTS

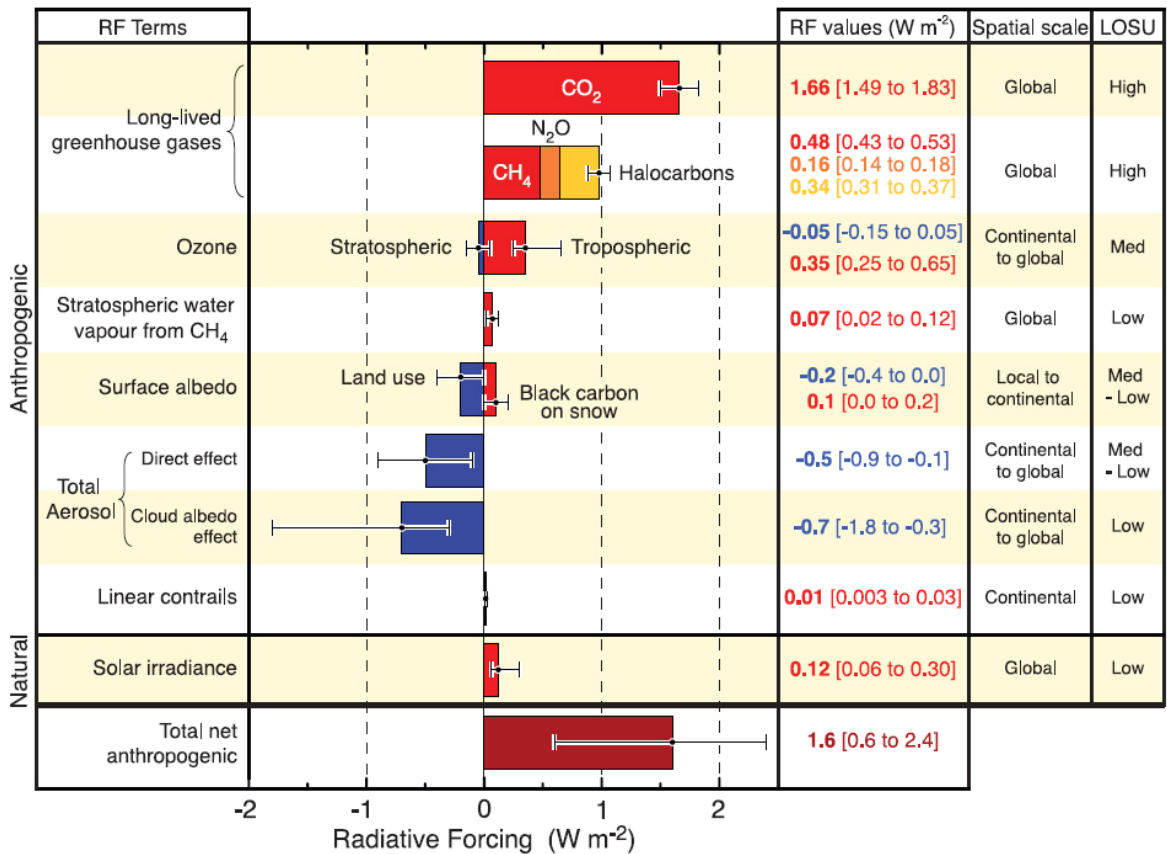


Fig 6- Radiative forcings for different anthropogenic and natural perturbations from 2005 relative to 1750. Re-printed from IPCC 2007.

The actual forcing over the last century has a wide range of uncertainty, largely due to the uncertain effect of aerosols (Fig. 6). Aerosols act to cool the surface by directly reflecting incoming sunlight back to space, but aerosol loads also influence clouds which have their own albedo response (the first indirect effect, or Twomey effect) and also the liquid water content, cloud height, and lifetime of clouds (the second indirect effect, or Albrecht effect). Aerosols have acted to offset an otherwise larger greenhouse warming over the last century, with the first indirect effect potentially being the largest (Koch et al., 2009; in press). Black Carbon can induce a slightly positive RF when deposited on snow, lowering the surface albedo, promote melting, and provide an overall warming effect.

As shown in Figure 2 and confirmed in other studies (e.g., Hansen et al., 2005) the Earth is currently taking in more radiation than it is emitting to space in the infrared. The current imbalance sustained over the last 10,000 years since the last glacial period could sufficiently raise the upper ocean temperature by over 100 °C. From this perspective, the current imbalance is relatively large and radiative equilibrium serves as a good approximation over long timescales. Indeed, the world's oceans

have undergone a significant increase in heat content and temperature since the industrial revolution (e.g., Levitus et al., 2005; Domingues et al., 2008).

4) *Climate Feedbacks*

In 1896, Svante Arrhenius calculated that a doubling of CO₂ in the atmosphere would produce a global temperature change of a little over 5 °C. This is likely an overestimate, but not by much. The first modern estimates were made over 30 years ago, and the same range of 1.5-4.5°C (now 2 to 4.5 °C in the latest IPCC report) has survived for decades, but unfortunately very little progress has come in narrowing this range. Narrowing this large range is very practical from a socio-economic view, and important scientifically for progression of our knowledge of climate change.

Climate feedbacks amplify or dampen forcings to the climate system. Feedbacks are net positive on timescales relevant for human consideration, so that the forcing factor will be amplified. Climate *sensitivity* refers to how responsive the system is to a given push. The terminology is typically used for discussions of global mean temperature, and typically does not consider the full range of non-temperature related responses, such as the possibility of abrupt climate shifts. The heat content of the land-atmosphere-ocean system is,

$$\frac{dH}{dt} = F - \frac{\Delta T}{\lambda} \quad 11$$

Where F is some perturbation of the climate system, and λ is the climate feedback parameter in (in W m⁻² °C⁻¹). The climate sensitivity parameter is the inverse (F/ΔT). In equilibrium, it follows that,

$$\Delta T = \lambda F \quad 12$$

The most powerful constraint working against a runaway temperature increase/decrease is the OLR, since radiation output is proportional to the fourth power of the temperature. This is usually already included in the response to a greenhouse perturbation, even in no-feedback calculations since it's a prerequisite for the planet to come back into equilibrium.

The predominant feedbacks in the climate system are water vapor changes, clouds, albedo responses, and the lapse rate. On a planet or satellite where a gas can condense into clouds (such as water on Earth, CO₂ on Mars, methane on Titan), increasing the temperature will increase the saturation vapor pressure, allowing more of that substance to build up before condensation. If that substance is a greenhouse gas, then increased concentration will provide a more enhanced greenhouse effect. Since water vapor is a very strong greenhouse gas, any warming will also increase atmospheric water vapor and provide a positive feedback. It turns out that water vapor provides the most powerful individual feedback, roughly doubling the sensitivity of climate to external changes (e.g., Bony et al., 2006).

Although water vapor's primary influence on the (enhanced) greenhouse effect is to reduce the OLR, it also absorbs a non-negligible amount of solar radiation. At the poles, this plays an important part in

feedback since there is a comparatively high chance of absorbing an upwelling photon in the visible wavelengths.

From Equation 7, it is clear that in an isothermal atmosphere where there is no lapse rate, then there can be no greenhouse effect and the average planetary temperature is that of the effective temperature. In a global warming situation, the tropical troposphere will warm faster than the surface as convection keeps the temperature profile close to the moist adiabat. Conversely, at the poles, the surface warming is amplified when compared to the atmosphere. The net effect globally is a small negative feedback as the lapse rate declines. The latent heat release of water vapor and the lapse rate changes are not independent, so the feedback is often grouped together (WV+LR, see Figure 7) with water vapor winning out.

The surface albedo response is generally the popular example given for a positive feedback in response to warming, since it is very intuitive and accurately depicts a cause → response → amplification relationship. The high Northern latitudes are currently warming at roughly twice the rate of the global mean (ACIA 2005). A primary cause is the ice-albedo effect in which the ratio of highly reflective ice to low reflective ocean/land declines, resulting in increased absorption of incoming solar radiation. The ice-albedo feedback effect declines in progressively warmer worlds (Colman and McAvaney 2009), until it is zero in an ice-free world. Changes in coverage of desert sand, forest area, etc also impact surface albedo responses. Contrary to popular impression, the ice-albedo effect in transient responses is much more pronounced in the Arctic than in Antarctica owing to Southern Ocean heat uptake and a higher fraction of land in the North, which is easier to heat than water (Holland and Bitz., 2003). Presumably the NH/SH responses would be more similar in equilibrium.

The ice-albedo effect is not as straightforward however as this simple explanation suggests. After all, the Arctic cold-season surface is also amplified when there is little to no incoming solar radiation. The albedo temperature effect is largest in the spring and autumn, and in summer, extra energy either goes into melting or into evaporation. However, heat absorbed by the ocean in summer is released back to the atmosphere; most of the heat release to the atmosphere comes from open water areas or areas of thin ice, so anomalous open water areas at the end of the melt season is to be associated with a positive surface air temperature anomaly in the cold months. As decades pass, there is more open water at the end of the melt season, and more heat in the Upper Ocean, and increased ocean-to-atmosphere transfer of heat in the cold season (Serreze et al., 2008). Other processes such as atmospheric transport patterns to the poles also play a role in Arctic amplification (Graversen et al., 2008).

Cloud feedbacks continue to be the most difficult problem in constraining climate sensitivity, even without taking into account the microphysical interaction between clouds and aerosols. In general, clouds have competing effects since they have a high albedo, but also reduce the OLR. Which effect wins out depends on cloud height, latitude, optical thickness, etc. The albedo response is dominated by low, cumulus clouds which are the largest source of uncertainty in feedback (e.g., Held and Soden 2006; Knutti and Hegerl 2008). High clouds, on the other hand, generally provide a net warming effect being relatively transparent in the shortwave, but very powerful in the infrared since they reside at very cold altitudes where they can have a large effect on the OLR. Two primary competing hypotheses regarding

high cloud responses are the IRIS hypothesis (Lindzen et al., 2001) and the Fixed Anvil Temperature (FAT) hypothesis (Hartmann and Larsson 2002).

The IRIS hypothesis proposes that precipitation efficiency of low-level clouds increases, resulting in decreased water detrained in the upper troposphere (a negative feedback since higher infrared-active clouds decline). The FAT hypothesis proposes that the temperature of high clouds tops remain unchanged and that their altitude increases with warming, a positive feedback. This has received support from high-resolution modeling. For example, Kuang and Hartmann 2007 show with high-resolution numerical modeling that anvil temperature changes less than a quarter of sea surface temperature (SST) changes, while the upper troposphere temperature change is larger than the SST change by a factor of three. In general, models tend to produce overall clouds feedbacks ranging from approximately neutral to strongly positive (e.g., Del Genio et al., 2005; Soden et al., 2008) and regional observations of decreases in low level cloud physical extent with temperature (e.g., Del Genio and Wolf 2000).

Again, at equilibrium,

$$Q_{in} = OLR(WV, C, GHG, T) \quad 13$$

Where the OLR is a function of the column distribution of greenhouse gases (GHG's) and water vapor (water vapor is a greenhouse gas, but will be separated since it is a feedback and not a forcing), clouds, and the vertical temperature profile. For our perturbation at the top-of-atmosphere,

$$\delta F = \delta F_T + \delta F_r + \delta F_{WV} + \delta F_{GHG} + \delta F_C + \delta F_\alpha \quad 14$$

For a no-feedback case where we perturb CO₂ and leave other variables constant (Held and Soden 2000),

$$\frac{\partial OLR}{\partial T} dT = - \frac{\partial OLR}{\partial \log_2 CO_2} d\log_2 CO_2 \quad 15$$

This becomes,

$$\frac{dT}{d\log_2 CO_2} = - \frac{\partial OLR}{\partial \log_2 CO_2} \bigg/ \frac{\partial OLR}{\partial T} \approx \frac{F}{4\sigma T_{eff}^3} \quad 16$$

Where $4\sigma T_{eff}^3$ comes from the slope of Stefan-Boltzmann. A doubling of CO₂ with no feedbacks raises the Earth's planetary temperature by about 1 °C. This equation can be further extended with feedbacks as the temperature and OLR change with water vapor, clouds, etc.

Under present Earth-like conditions, atmospheric water vapor increases nearly exponentially with temperature, with the saturation water vapor pressure e_s governed by the Clausius-Clapeyron equation (Pierrehumbert et al., 2007)

$$\begin{cases} e_s = e_s(T_0) \exp^{-5419(\frac{1}{T}-\frac{1}{273})} & \text{condensation to ice} \\ e_s = e_s(T_0) \exp^{-6148(\frac{1}{T}-\frac{1}{273})} & \text{condensation to liquid} \end{cases} \quad 17$$

Where T is some new temperature and 273 K is the baseline freezing temperature T_0 where $e_s(T_0) = 614$ pascals (1 pascal is .001 millibars). Models and observations suggest an approximately stable relative humidity in a climate change situation (e.g., Soden et al., 2005; Dessler et al., 2008), so under present temperatures we expect a roughly 7% increase in e_s per degree of warming.

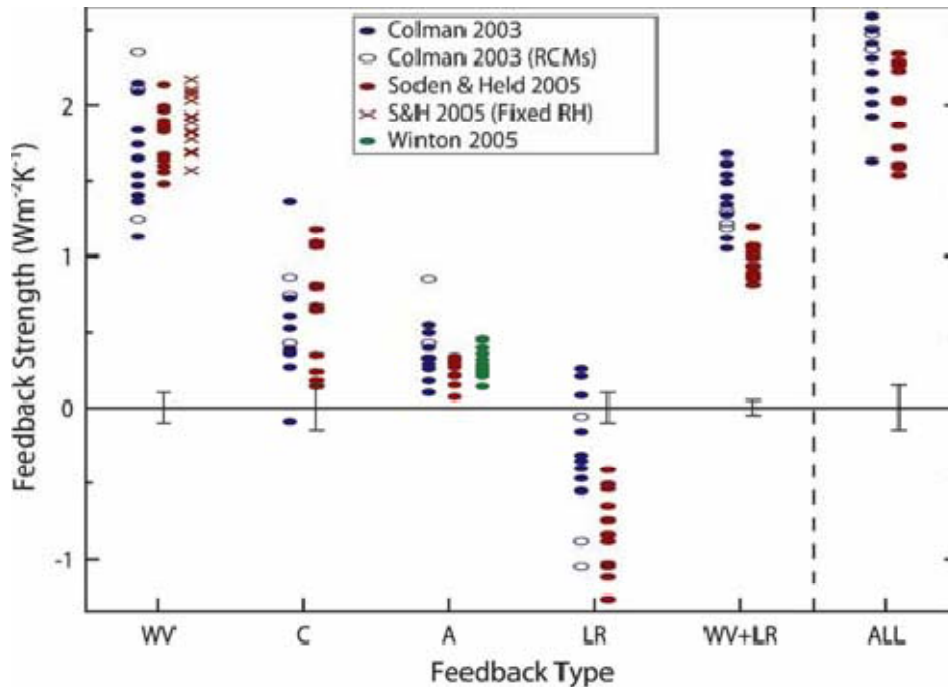


Fig 7- Feedback factor for various climate responses based on multiple studies. Re-printed from Bony et al., 2006

A reasonable question may be to ask how positive feedbacks can predominate without the system running away without bound. If each degree of warming leads to a feedback factor f of warming, this leads to f^2 of warming, etc., then,

$$\sum_{n=0}^{\infty} f^n = 1 + f + f^2 \dots = \frac{1}{1-f} \quad 18$$

Define the no-feedback climate sensitivity parameter as $\lambda_0 \approx 0.30 \text{ K/W m}^{-2} \text{ so,}$

$$\Delta T = \frac{F\lambda_0}{1-f} \quad 19$$

There are two obvious consequences to this. First, for some small uncertainty in f , larger uncertainties in the total climate sensitivity (and hence ΔT) are inevitable. Further, because climate sensitivity is proportional to $1/1-f$, it is difficult to rule out a very large sensitivity (Roe and Baker 2007) since the tail end on the large side increases abruptly (Fig. 8).

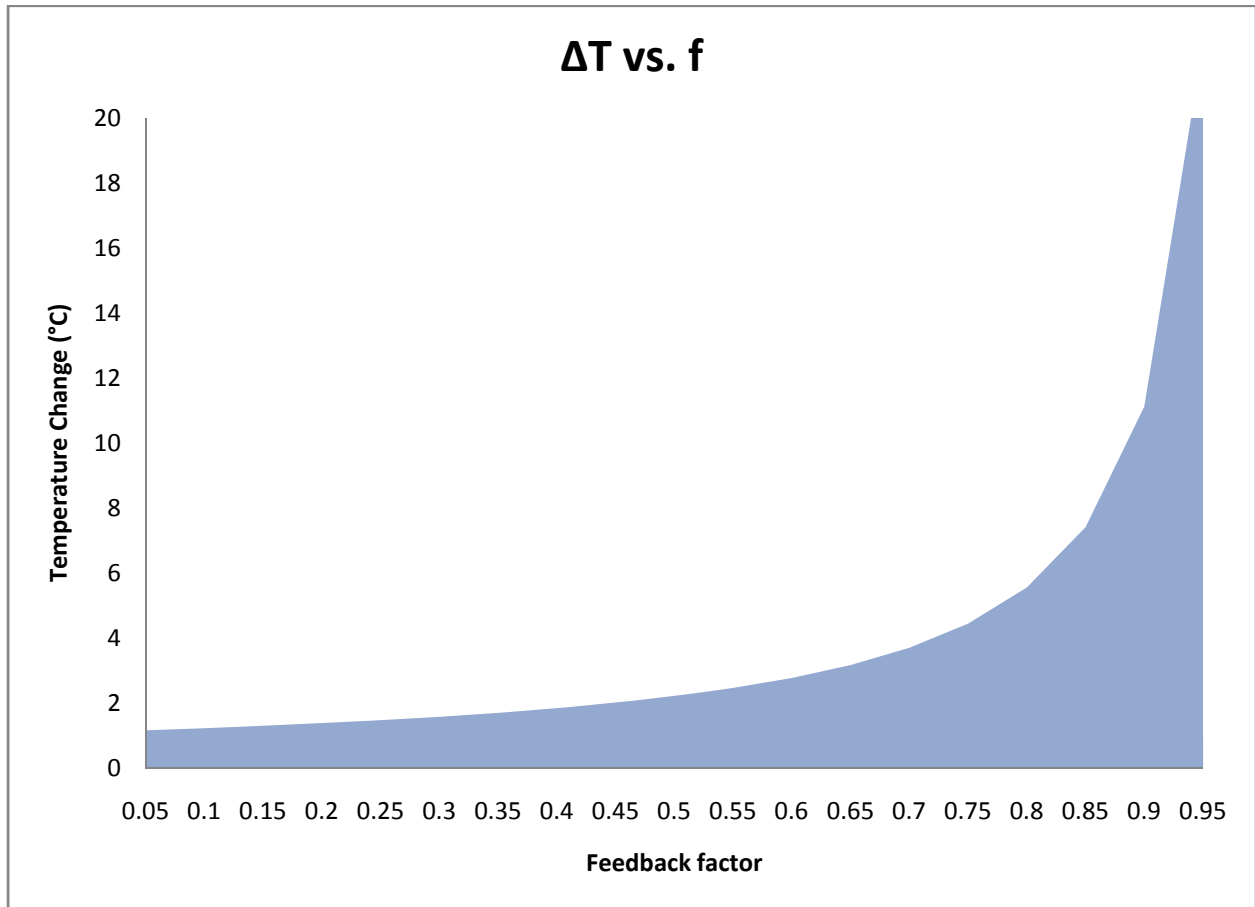


Fig 8- Changes in temperature (with a forcing equivalent to a doubling of carbon dioxide) as a function of the feedback factor

The second obvious constraint is that $0 < f < 1$ so as to produce a converging series (i.e., each feedback step is smaller than the last). If $f \geq 1$ then the result is a runaway greenhouse, which is clearly not physical for past or present Earth. For the currently accepted range of $2 \text{ °C} \leq \Delta T_{2 \times CO_2} \leq 4.5 \text{ °C}$ then $0.45 \leq f \leq 0.75$.

To illustrate the relationship and importance of individual feedbacks on each other, suppose for example that we introduce feedback factors f_{H_2O} and f_{α} for water vapor and ice/snow albedo effects. Then,

$$\frac{dT}{d\log_2 CO_2} \approx \frac{\Delta T_0}{1-f_{H_2O}-f_{\alpha}} \quad 20$$

Allow water vapor to be the only feedback gain to a doubling of CO_2 , for which f_{H_2O} is roughly 0.5 (NAS 2003). If $\Delta T_0 = 1$ °C then the total result would be a 2.0 °C change. However, suppose the albedo gain factor is 0.2. With just albedo changing, the total change would only be 1.25 °C. However, the combined effect would be for a temperature change of 3.3 °C. What is clear is that multiple feedbacks are related, since more water vapor provides an enhanced greenhouse effect that should melt more ice, which should in turn lower planetary albedo, increase temperature, and provide a further water vapor effect. Changes in cloud cover also impact surface albedo changes, since the impact of losing ice on the periphery of an ice sheet will be less with high cloud cover overhead.

5) *Constraining Climate Sensitivity*

There are multiple ways to put limits on climate sensitivity, which include paleoclimate constraints, attempting to use modern observations or individual events (e.g., volcanic eruptions), responses to the solar or seasonal cycles, and the use of Global Circulation Models (GCM's). When possible, theoretical tools must also be applied. GCM's must be tested against each other as well as past and present climate to see how robust the results are. Attempting to account for the full range of future climate evolution requires ensembles of multi-decadal simulations to assess internal climate variability as well as the model response uncertainty.

Attempting to use the 20th century observational record to constrain sensitivity is extremely difficult for two reasons. The first, which was discussed, is that the aerosol forcing (and thus the total forcing) is not well known. If the forcing is very small, then the sensitivity must be very large. If the forcing has been very large (e.g., essentially no negative aerosol influence), then the sensitivity may be smaller than modern estimates. The second problem is that the ice and ocean-atmosphere system has not yet equilibrated to the full change in greenhouse concentrations. Because of the immense thermal inertia of the oceans, it will take time until the full forcing is realized, typically on the order of decades. In fact, if all greenhouse emissions were to halt today, there would still be roughly half a degree of warming "in the pipeline" (e.g., Hansen et al., 2005) but this rate of ocean heat uptake has its own uncertainty. This amount of heating in the pipeline is also dependent on climate sensitivity, so if sensitivity is large, there would be a larger degree of warming that has yet to be realized than with a low-sensitive system.

Knutti et al., 2006 attempted to use the strength of the seasonal cycle and found a 5% chance that the sensitivity was below 1.5-2°C or above 5-6.5°C for a doubling of CO_2 , with a best estimate between 3 and 3.5 °C. Tung et al., 2008 attempted to use the well-defined forcing from the solar cycle and argued that the equilibrium sensitivity is larger than 5°C, but their analysis is relatively simple and likely underestimates thermal inertia.

The last glacial maximum (LGM, about 21 kyr) provides a very useful past example of a climate at equilibrium. Unfortunately, past data may have a wide range of uncertainty (either in constraining total

forcing for a particular time period, lack of global or hemispheric coverage, or in overall reconstruction of the climate variable in question). There is no past analogue for an abrupt increase in greenhouse gas concentrations, with all other factors influencing climate being the same as today.

Schneider von Deimling et al, 2006 test a multi-model ensemble of 1000 members against LGM data to show a similar sensitivity range of pre-IPCC 2007 values of 1.2–4.3 °C, with the possibility of an increased upper bound, and also conclude progress in reducing the range will involve including more regions with high-quality proxy-data for the LGM. Some authors have worked out climate sensitivity based on observed responses to volcanic eruptions, such as Mt. Pinatubo in 1991. A thorough discussion and analysis of observational-based work analyzing responses to the Mt. Pinatubo eruption, 20th century observations, as well as the LGM can be found in Annan and Hargreaves 2006.

The LGM is not the only example of paleoclimate constraint. Studies have analyzed forcings and responses over the full glacial-interglacial range or have used the last millennia to put constraints on sensitivity as well. Royer et al., 2007 interpret reconstructions from the past 420 million years and rule out a sensitivity < 1.5 °C where the largest ranges of their model parameters that fit proxy data well fall between 1.6 to 5.5 °C.

Even with more recent advances (see e.g., discussion in Knutti and Hegerl 2008) the likely range is now 2- 4.5°C and it is unlikely to lie outside that range.

6) Further Background

Global temperatures have risen roughly 0.8 °C over the last century (e.g., Brohan et al., 2006). Such numbers appear small with our daily intuition of weather fluctuations, but this is quite large when compared to global-scale, decadal-scale variability and is likely unusual in the context of the previous millennia and possibly much longer (e.g., Jones and Mann 2004; Moberg et al., 2005; Mann et al., 2008). For further perspective, the global range between today's relatively warm interglacial period and deep glacial times when ice covered much of North America and Europe is only about 5 °C with amplification at the poles. Additionally, the rate of present and 21st century climate change can exceed glacial-interglacial variations by an order of magnitude.

The equilibrium temperature response to a doubling of CO₂ would bring us into a new climate not witnessed since millions of years ago during the Mid-Pliocene (Dowsett et al., 2009) and the possibility of exceeding a doubling of pre-industrial CO₂ by the end of the century is not out of the question. During the Mid-Pliocene, sea levels were roughly 25 m higher and with substantial global ice loss relative to modern times. Although the response times to such sea level rise and ice sheet melt are not instantaneous, unperturbed emissions of fossil fuels (particularly a transition into a primary coal-based energy system) would produce extremely dangerous interference with the climate system with drastic ecological and socio-economic consequences (e.g., Hansen et al., 2008).

There is a vast amount of literature covering the current and projected impacts of climate change from the local to global scales. Such impacts include changes in frequency and/or intensity of droughts, heat waves, tropical storm activity, floods, and the like. Ecological responses, including poleward and upward (elevation) migration of plants, insects, and animals, changes in seasonal migration, egg-laying patterns, coral bleaching, etc have been detected at a global scale (e.g., Walther et al., 2002; Parmesan and Yohe 2003). Perhaps the most notable of observed 20th century response to climate change has been the rapid decline in the Arctic Sea Ice (ACIA 2005) whose extent is over 20% lower than values of just three decades ago, and predictions for a seasonally ice free Arctic ocean are on the timescales of a few decades. Extensive overviews of climate change impacts at the regional to global scales can be found in the Working Group 2 of IPCC 2007.

The extent of future climate changes not only includes uncertainties of climate sensitivity but on the trajectory of future emission scenarios. The evolution of socio-economic and energy policy is therefore crucial in evaluating the response to anthropogenic perturbation. CO₂ is rising at roughly 2 ppmv yr⁻¹ and following a 1% yr⁻¹ growth rate in emissions out to 100 years,

$$385 \text{ ppmv} + \int_0^{100} \left(\frac{2 \text{ ppmv}}{\text{yr}} \right) \exp^{0.01t} dt \approx 730 \text{ ppmv} \quad 21$$

In most cases, the release of enough gas or particle will quickly be mixed globally and produce forcing effects over a large distance, although certain particles (such as aerosols) tend to be more locally, regionally, and hemispherically constricted since its removal lifetime is often shorter than the time for global mixing.

The “extra pulse” of CO₂ that we release into the atmosphere will affect climate for centuries to millennia. Most of the excess that we release will be taken up by the oceans on a century timescales, but at least 20% will stay in the atmosphere for thousands of years as it reacts with igneous rocks and other slow removal processes (e.g., Archer and Brovkin 2008). The additional pulse of CO₂ will cause a temperature increase that will progress as the system tends toward equilibrium and be roughly constant for centuries even in the absence of further emissions. The world also does not end in 2100 or at a doubling of CO₂, so consideration for long-term trajectories is important. Matthews and Caldeira 2008 conclude that stabilizing global temperature at present-day levels requires emissions to be reduced to near-zero within a decade, which is virtually certain to be an unphysical scenario.

Conclusion

I have presented a simple description of radiative equilibrium, radiative forcings, and radiative feedback which determine global temperature and how it changes. The past record of climate changes shows that the sensitivity of climate is incompatible with a very low sensitivity, and probably a very high sensitivity as well. There is a greater chance of sensitivity being underestimated than overestimated, but decades of research with multiple methods currently point to best estimates between 2°C and 4.5

°C, with a central value of 3 °C. Stabilization at the current climate is virtually impossible given the degree of warming in the pipeline and atmospheric lifetime of extra CO₂. A relatively rapid curtailment of fossil-fuel emissions is necessary to avoid possible dangerous interference with the climate system.

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