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## Climate sensitivity in the Anthropocene

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531

### Abstract

Understanding the sensitivity of Earth's climate to an imposed external forcing is one of the great challenges in science and a critical component of efforts to avoid dangerous anthropogenic interference with the climate system. Climate sensitivity (or equilibrium global surface warming) to a doubling of atmospheric CO<sub>2</sub> has long been estimated to be about 3 °C, considering only fast climate feedbacks associated with increases in water vapor, decreases in sea ice, and changes in clouds. However, evidence from Earth's history suggests that slower surface albedo feedbacks due to vegetation change and melting of Greenland and Antarctica can come into play on the timescales of interest to humans, which could increase the sensitivity to significantly higher values, as much as 6 °C. Even higher sensitivity may result as present-day land and ocean carbon sinks begin to lose their ability to sequester anthropogenic CO<sub>2</sub> in the coming decades. The evolving view of climate sensitivity in the Anthropocene is therefore one in which a wider array of Earth system feedbacks are recognized as important. Since these feedbacks are overwhelmingly positive, the sensitivity is likely to be higher than has traditionally been assumed.

### 1 Introduction

The concept of climate sensitivity lies at the heart of climate system science. In its most basic form, it refers to the equilibrium change in global annual mean surface temperature that occurs in response to a radiative forcing, or externally imposed perturbation of the planetary energy balance. Typical forcings include variations in solar irradiance, atmospheric composition (e.g. due to natural processes such as volcanic eruptions, or human activities such as fossil fuel burning), and surface properties (e.g. due to anthropogenic land use change). Climate sensitivity depends critically on the sign and strength of climate feedbacks. A climate feedback is a change in Earth system properties which is induced by a climate forcing and which acts to either reinforce (for a

532



distribution/structure and the exposure of continental margins through changes in sea level). One can then define the fast (slow) feedback climate sensitivity as the particular case in which only fast (both fast and slow) feedback processes act to modify the basic Planck response to a forcing. (Note that the slow feedback climate sensitivity has been referred to elsewhere (e.g. Lunt et al., 2010) as the Earth system sensitivity.) In both cases, the sensitivity to specified changes in atmospheric GHG concentrations is considered. Thus, any carbon cycle feedbacks affecting atmospheric composition, which generally occur slowly over decades or longer, are included as part of the forcing (see Fig. 1a).

The classic fast feedback climate sensitivity problem was defined by Charney (1979) who considered the response to a doubling of atmospheric CO<sub>2</sub>. It was concluded, based largely on general circulation model (GCM) results, that the sensitivity is likely to lie between 1.5 °C and 4.5 °C, with a most probable value near 3 °C. Since the Charney report, a host of additional GCM and observational studies have attempted to estimate climate sensitivity based on the response to individual volcanic eruptions, climate change during the instrumental period (i.e. the last ~ 150 yr) and last millennium, Pleistocene glacial-interglacial transitions (e.g. from the last glacial maximum (LGM, ~20 thousand years (kyr) before present (BP)) to pre-industrial Holocene), and climate change occurring on longer timescales such as the Cenozoic (the past 65.5 million years (Myr)) and even the Phanerozoic (the past 545 Myr). Combining evidence from these studies suggests a most likely value and uncertainty range for the sensitivity similar to those given by Charney, but with higher sensitivities difficult to rule out (Hegerl et al., 2007; Knutti and Hegerl, 2008). It must be stressed, however, that this is the *fast feedback* climate sensitivity. Hansen et al. (2008) show that including slower surface albedo feedbacks associated with changes in continental ice sheets and vegetation increases the climate sensitivity for doubled CO<sub>2</sub> to about 6 °C for the range of climate states between glacial conditions and ice-free Earth. Other studies (Lunt et al., 2010) have found somewhat smaller values (~4–4.5 °C) for the slow feedback climate sensitivity. If the ice sheet/vegetation feedback is indeed this large, why has it not

535

received greater consideration? Most current GCMs are not equipped with interactive continental ice sheets. Any ice sheet changes occurring in these models (e.g. in LGM simulations) are therefore prescribed and are a forcing in the models. Similarly, empirical studies of climate sensitivity have typically regarded past ice sheet changes as an external forcing (see Fig. 1a). On one level this is to facilitate comparison with model results. More fundamentally, though, it is based on the long-standing notion that ice sheet changes occur so slowly (over several millennia) as to make them largely irrelevant to anthropogenic climate change occurring on timescales of decades to centuries. However, evidence from the paleoclimatic record for sea level changes of several meters per century (Thompson and Goldstein, 2005; Hearty et al., 2007), as well as present-day observations of increasing melt and overall mass loss from Greenland and Antarctica (Tedesco, 2007; Rignot and Jacobs, 2002; Zwally et al., 2002; Chen et al., 2006), suggest that ice sheet changes can occur more rapidly than previously assumed. This implies that the slow feedback climate sensitivity has relevance in the Anthropocene era (Hansen et al., 2008; Lunt et al., 2010), since ice sheet/vegetation feedback may become significant on decadal-to-centennial timescales of interest to humans.

Continued investigation is needed, however, in order to better constrain the range of possible magnitudes and the time dependence of the slow climate feedbacks (e.g. through more careful reconstructions of glacial-interglacial ice sheet and vegetation changes). For instance, while the magnitude of atmospheric GHG changes was about the same between the last interglacial (~125 kyr BP) and LGM and between the LGM and pre-industrial Holocene, the magnitude of the accompanying global temperature change was greater during the former period, indicating stronger amplifying feedbacks at work. This is supported by a larger sea level change between the last interglacial and LGM, which suggests a stronger ice sheet feedback. Thus, the strength of the slow feedbacks can vary depending on the particular time period considered, with potentially important implications for climate sensitivity.

536

#### 4 Carbon cycle feedbacks

Understanding and predicting future climate change requires knowledge of the CO<sub>2</sub> sequestration capacity of the land and ocean and its changes over time, which determine how anthropogenic CO<sub>2</sub> emissions translate into changes in atmospheric CO<sub>2</sub> concentration. Only a portion of present-day emissions actually remains in the atmosphere. This “airborne fraction” varies somewhat from year-to-year, but was estimated to be 43 % on average between 1959 and 2008 (Le Quéré et al., 2009). The remaining portion (57 %) of CO<sub>2</sub> emissions during this time was taken up by the ocean and terrestrial biosphere. If the airborne fraction was to remain constant over time, it would be straightforward to calculate the emissions trajectory that would be required to achieve stabilization of atmospheric CO<sub>2</sub> at a given level. This is not expected to be the case, however, as the airborne fraction is anticipated to increase as a result of future climate change (Friedlingstein et al., 2006; Plattner et al., 2008; Archer et al., 2009). In other words, land and ocean carbon sinks will become less efficient at absorbing anthropogenic CO<sub>2</sub> if atmospheric CO<sub>2</sub> continues to rise rapidly, thus producing a positive feedback on atmospheric CO<sub>2</sub> levels and consequently climate warming. The strength of this feedback varies substantially between different coupled climate-carbon cycle models. By the end of the twenty-first century, these models predict an increase in atmospheric CO<sub>2</sub> of anywhere from 20 to 200 ppm as a result of climate-carbon cycle feedbacks, which leads to an additional climate warming of between 0.1 and 1.5 °C (Friedlingstein et al., 2006). While the representation of several processes in the current generation of climate-carbon cycle models is uncertain, it is important to point out that the models are in qualitative agreement with paleodata from the Pleistocene (see supplementary information).

Land and ocean carbon sinks are expected to be affected both by the increase in atmospheric CO<sub>2</sub> itself, and by the climate changes resulting from this CO<sub>2</sub> increase. Over land, higher atmospheric CO<sub>2</sub> levels are likely to have some stimulatory effect on plant photosynthesis which would have a negative impact on the CO<sub>2</sub> growth rate.

537

The strength of this CO<sub>2</sub> fertilization effect, however, particularly in the long term, is unclear and depends critically on the availability of reactive nitrogen (Denman et al., 2007; Hyonen et al., 2007; Reich et al., 2006; Heimann and Reichstein, 2008; Gruber and Galloway, 2008; Zaehle et al., 2010; Arneeth et al., 2010). Higher temperatures will impact both net primary production (NPP, the difference between photosynthesis and autotrophic respiration) and heterotrophic respiration  $R_h$ , which (along with disturbance such as wild fire and land-use change) determine the net carbon balance of terrestrial ecosystems. NPP is expected to generally increase at high latitudes due to extended growing seasons.  $R_h$  is typically assumed to increase with temperature, although the magnitude and time dependence of this effect are debated (Giardina and Ryan, 2000; Luo et al., 2001; Knorr et al., 2005; Kirschbaum, 2004; Davidson and Janssens, 2006). Other climate changes, in particular changes in the hydrological cycle, will also affect NPP and  $R_h$ , yet these changes are often more uncertain and more regionally variable and model dependent than temperature changes. It is also important to consider how changes in anthropogenic land use and management may impact terrestrial ecosystem carbon balance. At present, 32 % of the global ice free land surface is used for agriculture (Foley et al., 2007), and almost 25 % of the global potential NPP is appropriated directly and indirectly by humans (Haberl et al., 2007). Increasing population and needs for food will significantly change the future dynamics of the land carbon sink.

The uptake of atmospheric CO<sub>2</sub> by the ocean depends on the difference in CO<sub>2</sub> partial pressure ( $p\text{CO}_2$ ) between the air and surface water. Surface water  $p\text{CO}_2$  is regulated by the series of chemical reactions that describe the ocean’s carbonate system. When CO<sub>2</sub> is added to seawater, the net effect is a reaction with carbonate ion to form bicarbonate ion, which reduces the amount of carbonate available to react with further CO<sub>2</sub> additions. This increases the  $p\text{CO}_2$  of the seawater and thus decreases the ocean’s “buffering capacity” to draw down atmospheric CO<sub>2</sub> (Denman et al., 2007). Atmospheric CO<sub>2</sub> uptake is also determined by the rate of ocean vertical mixing. Most GCMs suggest that global warming will be accompanied by a weakening of the ocean’s thermohaline circulation and associated reduction in the rate of mixing

538

between surface and deep waters (Meehl et al., 2007), which would tend to reduce CO<sub>2</sub> uptake by decreasing the effective volume of the ocean that is exposed to the atmosphere. Changes in ocean vertical mixing as well as temperature and pH would also affect the biological component of the ocean's carbon cycle (Sarmiento et al., 2004), which would have further implications for the uptake of anthropogenic CO<sub>2</sub>. In summary, although the carbon cycle is clearly complex and several key processes are still incompletely understood, there is the expectation that the present-day land and ocean sinks for anthropogenic CO<sub>2</sub> will weaken in the coming decades as climate change progresses.

It is also important to consider possible changes in the sources and sinks of other GHG besides CO<sub>2</sub>. For example, atmospheric methane (CH<sub>4</sub>) variations are known to have closely tracked global temperature changes throughout Earth's climatic past (Chappellaz et al., 1993; Beerling et al., 2009), and increases in CH<sub>4</sub> during the industrial era produced the second-largest radiative forcing of the well-mixed GHG after CO<sub>2</sub> (Forster et al., 2007). CH<sub>4</sub> has a much stronger infrared absorption capacity than CO<sub>2</sub> on a per molecule basis, and has a higher "efficacy" than CO<sub>2</sub> due mainly to its tendency to increase tropospheric ozone and stratospheric water vapor (Hansen et al., 2005b). The dominant natural source of atmospheric CH<sub>4</sub> is emissions from continental wetlands (Bartlett and Harriss, 1993), implying that CH<sub>4</sub>-climate feedbacks will depend strongly on future changes in the hydrological cycle. For example, projected increases in high latitude precipitation (Meehl et al., 2007) could increase CH<sub>4</sub> emissions from northern peatlands, which would contribute to climate warming. However, peatlands currently remove CO<sub>2</sub> from the atmosphere during photosynthesis and sequester a portion of the carbon in accumulating peat (Frolking et al., 2006; Frolking and Roulet, 2007), and thus any changes in CO<sub>2</sub> sequestration must also be factored in when determining net climate impact. It is generally expected that changes in CH<sub>4</sub> emissions may be important on decadal timescales, but that on century-to-millennial timescales CO<sub>2</sub> effects will dominate as a result of the much longer time required for atmospheric CO<sub>2</sub> to reach a new equilibrium following a perturbation to the peatland

539

carbon sink (Frolking et al., 2006). Other natural sources of atmospheric CH<sub>4</sub>, though relatively small at present, could become important in the future. For instance, destabilization of methane clathrates on the ocean floor caused by higher temperatures could trigger the release of CH<sub>4</sub> into the atmosphere which would amplify global warming. This mechanism has been proposed (Brook et al., 2008) to explain the massive carbon release and pronounced warming that occurred during the Paleocene-Eocene Thermal Maximum (PETM, ~55 Myr BP).

Changes in the nitrogen cycle are another important consideration. Human actions through food and energy production have profoundly altered the abundance and availability of reactive N on the Earth's surface (Galloway et al., 2008). In addition to a number of other impacts, nitrogen species have both direct and indirect impacts on climate change and as such, possible changes in their sources and sinks will affect the magnitude of those impacts. The direct effects are associated with nitrous oxide (N<sub>2</sub>O), ozone and N-containing aerosols. The first two are GHG and their increased abundance (in the troposphere for ozone) due to human activity has produced a positive forcing (warming); the third contributes to negative forcing (cooling) (Forster et al., 2007). Indirect impacts are through C-N interactions in ecosystems, both terrestrial (Gruber and Galloway, 2008) and marine (Duce et al., 2008). It is likely that these impacts will increase with time due to population growth, and increased per-capita use of agricultural resources (Erisman et al., 2008).

## 5 Implications and ways forward

Our present understanding of climate sensitivity is the product of extensive research utilizing both climate models and observations and focusing on diverse periods throughout Earth's history. And if past is prologue, climate sensitivity during the current Anthropocene era is likely to be high due to the dominance of amplifying (positive) feedbacks on decadal-to-centennial timescales of interest to humans. Fast feedbacks associated with changes in atmospheric temperature, water vapor, clouds, sea ice, and snow

540

cover amplify the basic Planck response to forcing by a factor of  $\sim 1.5$ – $4.5$ , with a best estimate for the fast feedback climate sensitivity for doubled  $\text{CO}_2$  of about  $3^\circ\text{C}$ . Additionally, there is increasing evidence that slower surface albedo feedbacks associated with changes in the area of continental ice sheets may become important during the Anthropocene, which could increase the  $2\times \text{CO}_2$  sensitivity to as much as  $6^\circ\text{C}$ .  
5 A climate sensitivity of  $6^\circ\text{C}$  for doubled  $\text{CO}_2$  ( $\Delta F = 3.7 \text{ W m}^{-2}$ ) would imply a climate feedback parameter  $\lambda$  of  $0.6 \text{ W m}^{-2} \text{ }^\circ\text{C}^{-1}$ , indicating that an additional  $1.4^\circ\text{C}$  of global warming is still “in the pipeline” as a result of past forcing not yet responded to (i.e. the present-day planetary energy imbalance of  $0.85 \text{ W m}^{-2}$ ). This committed warming is on top of the  $\sim 0.8^\circ\text{C}$  warming that has already occurred (Hansen et al., 2010),  
10 bringing global temperatures to about  $2.2^\circ\text{C}$  above pre-industrial levels. Thus, if the goal is to limit global warming to below  $2^\circ\text{C}$  (which is commonly used as the threshold beyond which dangerous climate change could occur; e.g. European Council, 2005), a  $6^\circ\text{C}$  climate sensitivity would signify that the current atmospheric  $\text{CO}_2$  concentration  
15 of  $\sim 390$  ppm may already be in the danger zone. This is further supported by the recent finding that global sea level during the Middle Pliocene (3.0–3.5 Myr BP), a time with atmospheric  $\text{CO}_2$  levels similar to today, was about 25 m higher than at present (Rohling et al., 2009). Finally, it is worth noting that the idea of committed change due to past forcing may be applicable not just to the response of the physical system (e.g. global temperature and sea level change), but to the response of ecosystems as well  
20 (Jones et al., 2009).

Narrowing the range of uncertainty in climate sensitivity estimates has proven to be an arduous task (Knutti and Hegerl, 2008). However, future efforts toward this end will benefit from improved characterization of climate forcing and response, such as more  
25 accurate and spatially and temporally complete paleo-reconstructions, and more tightly constrained estimates of aerosol forcing and ocean heat uptake during the instrumental era. Additionally, there is the need for better theoretical understanding of several key feedback processes. For example, although not discussed here, cloud feedback has long been identified as the primary source of uncertainty in climate sensitivity in

models, yet an adequate understanding of the factors controlling the global cloud distribution and how these may change with global temperature continues to be lacking. Loss of significant ice mass from Greenland and Antarctica could produce important  
5 feedbacks on future climate change, yet these feedbacks are likely to depend critically on dynamical processes such as ice stream acceleration and ice shelf disintegration that are not represented in current ice sheet models (Dupont and Alley, 2006). Ice sheet melting may be important not just for its effect on surface albedo, but also because of other feedbacks it may induce such as changes in the ocean’s thermohaline circulation (Swingedouw et al., 2008; Goelzer et al., 2010). Carbon cycle models  
10 generally neglect nutrient (e.g. nitrogen, phosphorus, iron) limitations on primary production, peatland/permafrost dynamics, changes in fire frequency, the response of marine ecosystems to ocean acidification, and other potentially important biogeochemical processes, thus hampering our ability to quantify climate-carbon cycle feedbacks on decadal-to-centennial timescales.

15 What is the conceptual framework for climate sensitivity best suited for the Anthropocene? In other words, how should the traditional paradigm be expanded based on what we presently know about Earth system feedbacks that may be relevant to humans? Figure 1a and b illustrates how our understanding of climate sensitivity has evolved. The traditional view of continental ice sheets as slowly varying or fixed boundary  
20 forcings has been replaced by one in which ice sheets can respond to and then amplify anthropogenic global warming. Similarly, there is the need to view land and ocean carbon sinks as interactive components of the system capable of producing important climate feedbacks. With the development of Earth system models and future increases in computing power, it should be possible in time to estimate climate sensitivity  
25 including ice sheet and carbon cycle feedbacks. Such estimates can already be obtained from empirical data by changing what we classify as forcing versus feedback. We conclude that since ice sheet and carbon cycle feedbacks are both expected to be positive on human relevant timescales, climate sensitivity in the Anthropocene is likely to be higher than the fast feedback sensitivity that has typically been assumed (Hansen

et al., 2008; Lunt et al., 2010; Pagani et al., 2010; Kiehl, 2011; Park and Royer, 2011). However, it is important to remember that humans, like other *internal* components of the Earth system (see Fig. 1b), are capable of responding to ongoing climate change (and, unlike these other components, can also respond to the *anticipation* of climate change). Therefore, any positive or negative feedbacks associated with changes in human behavior (e.g. changes in fossil fuel burning, land use and land/ocean ecosystem management) will also be important for climate sensitivity. Such anthropogenic feedbacks represent perhaps the greatest source of uncertainty in future climate change projections.

**Supplementary material related to this article is available online at:**  
<http://www.earth-syst-dynam-discuss.net/2/531/2011/esdd-2-531-2011-supplement.pdf>.

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543

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544

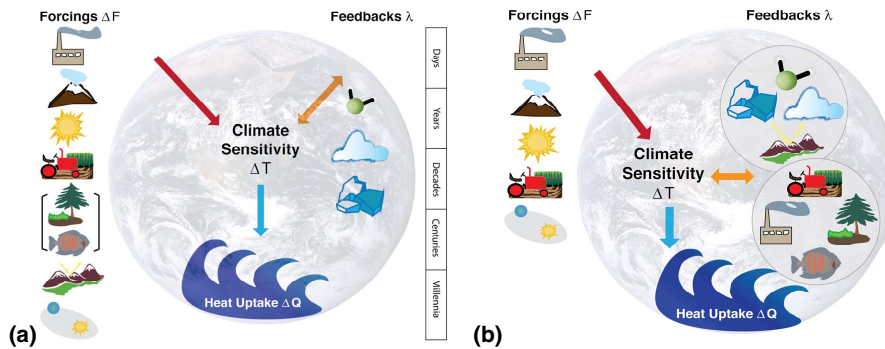




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**Fig. 1.** A climate forcing  $\Delta F$  triggers a series of feedbacks (represented by the feedback parameter  $\lambda$ ) which determine the resulting equilibrium global mean surface temperature change, or climate sensitivity,  $\Delta T$ . Delay in this equilibrium temperature response due to ocean and cryosphere inertia leads to a net planetary heat uptake  $\Delta Q$ . **(a)** Traditional framework: Climate sensitivity to an imposed external forcing depends solely on fast climate feedbacks occurring on timescales of decade/s or less, specifically changes in water vapor, clouds, and sea ice. Processes regarded as forcings are (from top to bottom) anthropogenic perturbations of atmospheric composition (including greenhouse gases and aerosols) due to fossil fuel burning, volcanic eruptions, variations in solar luminosity, changes in anthropogenic land use and land/ocean ecosystem management, changes in terrestrial carbon sequestration, changes in ocean carbon sequestration, surface albedo changes from land ice and vegetation, and variations in insolation (incoming solar radiation) due to changes in Earth's orbit. **(b)** New framework: Several Earth system processes traditionally regarded as external forcings are now considered to be internal feedbacks contributing to climate sensitivity, in particular surface albedo changes associated with the waxing and waning of continental ice sheets, changes in land and ocean carbon sinks, and the human response to ongoing and anticipated climate change. (Note that human behavior changes are both a forcing and a feedback, since they can initiate Earth system change and also be a response to that change.) Feedbacks are viewed as perturbations to the water and carbon cycles occurring across multiple timescales.