Earth Syst. Dynam. Discuss., 2, 531–550, 2011
www.earth-syst-dynam-discuss.net/2/531/2011/
doi:10.5194/esdd-2-531-2011

Earth System Dynamics Discussions

Climate sensitivity in the Anthropocene

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Received: 2 September 2011 – Accepted: 8 September 2011
– Published: 15 September 2011

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Published by Copernicus Publications on behalf of the European Geosciences Union.

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_2 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_2 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

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positive feedback) or counteract (for a negative feedback) the forcing. In the traditional framework, illustrated by Fig. 1a, climate sensitivity to an applied forcing is determined by fast climate feedbacks occurring on timescales of decade/s or less, specifically changes in water vapor, clouds, and sea ice. Slower surface albedo feedbacks associated with changes in land ice (e.g. continental ice sheets, mountain glaciers) and vegetation are either not considered or are included as part of the forcing. Additionally, any changes in terrestrial and ocean carbon sequestration are implicit (as denoted by brackets in Fig. 1a) since changes in atmospheric greenhouse gas (GHG) concentrations are specified as forcing. While the fast feedback climate sensitivity has long been the accepted paradigm, there is increasing evidence that slower amplifying ice sheet and carbon cycle feedbacks may become important on decadal-to-centennial timescales of interest to humans. This indicates the need to redefine the traditional framework to explicitly account for these (and potentially other) feedbacks, and suggests that climate sensitivity in the Anthropocene (Crutzen and Stoermer, 2000; Zalasiewicz et al., 2008) is likely to be higher than previously assumed.

2 Earth's energy balance

In response to a positive radiative forcing ΔF, such as characterizes the present-day anthropogenic perturbation (Forster et al., 2007), the planet must increase its longwave (LW) emission to space in order to re-establish energy balance (see supplementary information). Assuming that this increased LW emission is proportional to the surface temperature change ΔT, we can write

$$\Delta F = \lambda \Delta T + \Delta Q \quad (1)$$

where λ is the climate feedback parameter. Complete restoration of the planetary energy balance (and thus full adjustment of the surface temperature) does not occur instantaneously due to the inherent inertia of the system, which lies mainly in the slow response times of the oceans and cryosphere. Therefore, prior to achieving a

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new equilibrium state, there will be an imbalance, ΔQ, between climate forcing and response. This imbalance represents the net heat flux into the system, with nearly all of this heat flux going into the ocean (Levitus et al., 2005). For long-lived forcings such as increases in well-mixed GHG, exchange of heat between the upper mixed layer and deep ocean can delay the full surface temperature response by decades to centuries, with this delay also being a strong (quadratic) function of climate sensitivity (Hansen et al., 1985). At present, ΔQ (referred to herein as the ocean heat uptake) is estimated to be $0.85 \pm 0.15\,\mathrm{W\,m^{-2}}$ (Hansen et al., 2005a), implying that additional global warming is still "in the pipeline" even without any further changes in radiative forcing.

For a given forcing ΔF, λ is determined by two factors: the basic Planck (or black-body) response of the Earth's LW emission that is required to balance the forcing, and any feedbacks that come into play as the planet warms. It is readily shown that for present-day Earth, the Planck response is $\lambda_0 \approx 3.8\,\mathrm{W\,m^{-2}\,{}^{\circ}C^{-1}}$ (see supplementary information). Therefore, *in the absence of any feedbacks* (i.e. $\lambda = \lambda_0$), a doubling of the atmospheric carbon dioxide (CO_2) concentration, which represents a forcing $\Delta F = 3.7\,\mathrm{W\,m^{-2}}$ (Forster et al., 2007), would produce an equilibrium ($\Delta Q = 0$) surface warming of about $1\,{}^{\circ}\mathrm{C}$. As will be discussed, however, the true equilibrium climate sensitivity is expected to be larger than this, perhaps substantially so, as a result of strong amplifying (positive) feedbacks operating within the Earth's system.

3 Fast versus slow feedback sensitivity

Climate feedbacks depend on the timescale considered, the characteristics of the forcing (e.g. spatial pattern, spectral dependence), and the climate state when the forcing is applied. In this section we focus on the timescale dependence of climate feedbacks. Fast feedbacks occurring on timescales of decade/s or less are associated with changes in atmospheric temperature, water vapor, clouds, sea ice, and snow cover. Slow surface albedo feedbacks occurring over decades or longer are associated with the waxing and waning of continental ice sheets (and related effects on vegetation

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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_2. 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_2 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

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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.

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4 Carbon cycle feedbacks

Understanding and predicting future climate change requires knowledge of the CO_2 sequestration capacity of the land and ocean and its changes over time, which determine how anthropogenic CO_2 emissions translate into changes in atmospheric CO_2 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_2 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_2 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_2 if atmospheric CO_2 continues to rise rapidly, thus producing a positive feedback on atmospheric CO_2 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_2 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_2 itself, and by the climate changes resulting from this CO_2 increase. Over land, higher atmospheric CO_2 levels are likely to have some stimulatory effect on plant photosynthesis which would have a negative impact on the CO_2 growth rate.

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The strength of this CO_2 fertilization effect, however, particularly in the long term, is unclear and depends critically on the availability of reactive nitrogen (Denman et al., 2007; Hyvonen et al., 2007; Reich et al., 2006; Heimann and Reichstein, 2008; Gruber and Galloway, 2008; Zaehle et al., 2010; Arneth 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_2 by the ocean depends on the difference in CO_2 partial pressure (pCO_2) between the air and surface water. Surface water pCO_2 is regulated by the series of chemical reactions that describe the ocean's carbonate system. When CO_2 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_2 additions. This increases the pCO_2 of the seawater and thus decreases the ocean's "buffering capacity" to draw down atmospheric CO_2 (Denman et al., 2007). Atmospheric CO_2 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

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between surface and deep waters (Meehl et al., 2007), which would tend to reduce CO_2 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_2. 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_2 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_2. For example, atmospheric methane (CH_4) 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_4 during the industrial era produced the second-largest radiative forcing of the well-mixed GHG after CO_2 (Forster et al., 2007). CH_4 has a much stronger infrared absorption capacity than CO_2 on a per molecule basis, and has a higher "efficacy" than CO_2 due mainly to its tendency to increase tropospheric ozone and stratospheric water vapor (Hansen et al., 2005b). The dominant natural source of atmospheric CH_4 is emissions from continental wetlands (Bartlett and Harriss, 1993), implying that CH_4-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_4 emissions from northern peatlands, which would contribute to climate warming. However, peatlands currently remove CO_2 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_2 sequestration must also be factored in when determining net climate impact. It is generally expected that changes in CH_4 emissions may be important on decadal timescales, but that on century-to-millennial timescales CO_2 effects will dominate as a result of the much longer time required for atmospheric CO_2 to reach a new equilibrium following a perturbation to the peatland

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carbon sink (Frolking et al., 2006). Other natural sources of atmospheric CH_4, 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_4 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_2O), 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

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cover amplify the basic Planck response to forcing by a factor of ~1.5–4.5, with a best estimate for the fast feedback climate sensitivity for doubled CO_2 of about 3 °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× CO_2 sensitivity to as much as 6 °C. A climate sensitivity of 6 °C for doubled CO_2 ($\Delta F = 3.7\,\mathrm{W\,m^{-2}}$) would imply a climate feedback parameter λ of $0.6\,\mathrm{W\,m^{-2}\,°C^{-1}}$, indicating that an additional 1.4 °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 $0.85\,\mathrm{W\,m^{-2}}$). This committed warming is on top of the ~0.8 °C warming that has already occurred (Hansen et al., 2010), bringing global temperatures to about 2.2 °C above pre-industrial levels. Thus, if the goal is to limit global warming to below 2 °C (which is commonly used as the threshold beyond which dangerous climate change could occur; e.g. European Council, 2005), a 6 °C climate sensitivity would signify that the current atmospheric CO_2 concentration of ~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 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 (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 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

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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 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 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.

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 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 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

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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.

Acknowledgements. We thank Reto Knutti for helpful comments on the manuscript. This work was funded by a grant from the LDEO/GISS Climate Center, and was motivated by discussions that took place at a meeting titled "Climate Sensitivity Extremes: Assessing the Risk" that was held at NASA GISS during April 2010.

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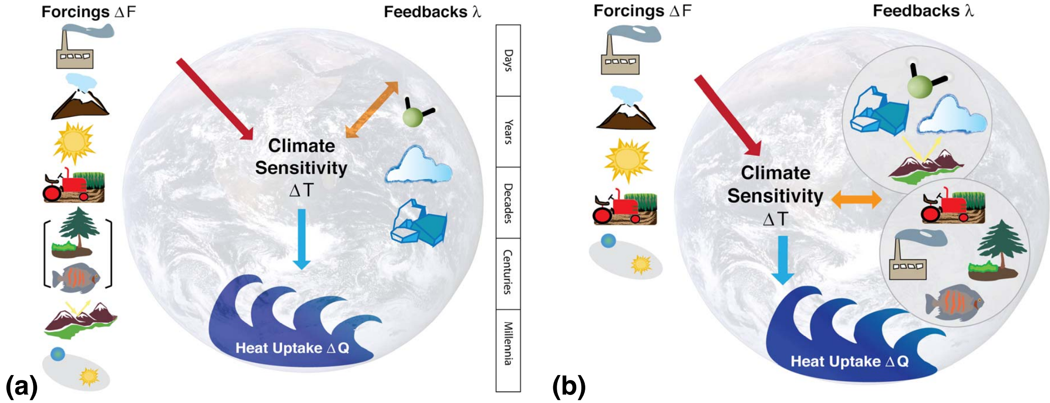

Fig. 1. A climate forcing ΔF triggers a series of feedbacks (represented by the feedback parameter λ) which determine the resulting equilibrium global mean surface temperature change, or climate sensitivity, ΔT. Delay in this equilibrium temperature response due to ocean and cryosphere inertia leads to a net planetary heat uptake ΔQ. **(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.

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