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Review Article
Climate sensitivity in the Anthropocene

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Climate sensitivity in its most basic form is defined as the equilibrium change in global surface temperature that occurs in response to a climate forcing, or externally imposed perturbation of the planetary energy balance. Within this general definition, several specific forms of climate sensitivity exist that differ in terms of the types of climate feedbacks they include. Based on evidence from Earth's history, we suggest here that the relevant form of climate sensitivity in the Anthropocene (e.g. from which to base future greenhouse gas (GHG) stabilization targets) is the Earth system sensitivity including fast feedbacks from changes in water vapour, natural aerosols, clouds and sea ice, slower surface albedo feedbacks from changes in continental ice sheets and vegetation, and climate–GHG feedbacks from changes in natural (land and ocean) carbon sinks. Traditionally, only fast feedbacks have been considered (with the other feedbacks either ignored or treated as forcing), which has led to estimates of the climate sensitivity for doubled CO_2 concentrations of about 3°C. The $2\times CO_2$ Earth system sensitivity is higher than this, being ∼4–6°C if the ice sheet/vegetation albedo feedback is included in addition to the fast feedbacks, and higher still if climate–GHG feedbacks are also included. The inclusion of climate–GHG feedbacks due to changes in the natural carbon sinks has the advantage of more directly linking anthropogenic GHG emissions with the ensuing global temperature increase, thus providing a truer indication of the climate sensitivity to human perturbations. The Earth system climate sensitivity is difficult to quantify due to the lack of palaeo-analogues for the present-day anthropogenic forcing, and the fact that ice sheet and climate–GHG feedbacks have yet to become globally significant in the Anthropocene. Furthermore, current models are unable to adequately simulate the physics of ice sheet decay and certain aspects of the natural carbon and nitrogen cycles. Obtaining quantitative estimates of the Earth system sensitivity is therefore a high priority for future work.

Key Words: climate sensitivity; radiative forcing; radiative feedbacks; carbon cycle; Anthropocene

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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. Within this general definition, however, there exist several specific forms of the climate sensitivity. It is important to distinguish between these forms in order to avoid confusion and to reconcile results from different studies that employ alternative sensitivity definitions. The goals of this article are therefore to clarify the various meanings of climate sensitivity, and to suggest the form of the sensitivity that is most relevant in the Anthropocene era (Crutzen and Stoermer, 2000; Zalasiewicz *et al.*, 2008).

Climate sensitivity definitions differ in terms of the types of climate feedbacks that they include (Figure 1). A climate feedback is an Earth system response to a climate forcing that either reinforces (for a positive feedback) or counteracts (for a negative feedback) the forcing. We consider here three main types of climate sensitivity: (i) the fast feedback sensitivity (Figure 1(a)), (ii) the Earth system sensitivity including ice sheet and vegetation albedo feedbacks (Figure 1(b)), and (iii) the Earth system sensitivity additionally including climate–greenhouse gas (GHG) feedbacks (Figure 1(c)).

The traditional and most widely used form of the climate sensitivity is the fast feedback sensitivity (Figure 1(a)). In this case, climate sensitivity to an applied forcing is determined solely by fast climate feedbacks occurring on time-scales of decade(s) or less[†], specifically changes in water vapour, natural aerosols, 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 part of the forcing. Additionally, no attempt is made to discriminate between changes in atmospheric GHG concentrations due to anthropogenic emissions and those due to changes in the natural carbon sinks, since the forcing is regarded as the *total* atmospheric GHG change. (The canonical forcing is a doubling of the atmospheric CO_2 concentration.) Thus, any changes in terrestrial and ocean carbon sequestration are implicit, as denoted by brackets in Figure 1(a).

While the fast feedback sensitivity has long been the accepted paradigm (e.g. in assessment reports of the Intergovernmental Panel on Climate Change (IPCC)), there is mounting evidence to suggest that additional feedbacks should be included in the definition of climate sensitivity to GHG forcing. Ice sheet and vegetation albedo feedbacks occur relatively slowly over centuries or longer, yet they have the potential to become significant in the Anthropocene due to the very long lifetime (centuries to millennia) of fossil fuel CO_2 (Archer *et al.*, 2009). Climate sensitivity with ice sheet and vegetation feedbacks included is typically referred to as the Earth system sensitivity (Figure 1(b)). It is greater than the fast feedback sensitivity since ice sheets melt as the climate warms, thereby decreasing the surface albedo and producing further warming. One can also consider a more comprehensive form of the Earth system sensitivity that additionally incorporates climate–GHG feedbacks (Figure 1(c)). These feedbacks are expected to be positive, since climate warming diminishes the ability of the oceans and terrestrial biosphere to sequester anthropogenic carbon. Including climate–GHG feedbacks in the Earth system sensitivity thus implies a higher sensitivity still.

The remainder of this article is organized as follows. Section 2 discusses climate sensitivity within the framework of Earth's energy balance. Following this, we describe in more detail the three types of climate sensitivity already alluded to, specifically the fast feedback sensitivity (section 3), the Earth system sensitivity including ice sheet and vegetation albedo feedbacks (section 4), and the Earth system sensitivity additionally including climate–GHG feedbacks (section 5). Finally, in sections 6 and 7, we present conclusions and discuss future directions in climate sensitivity research.

2. Earth's energy balance

In response to a positive radiative forcing ΔF (see Appendix A), such as characterizes the present-day anthropogenic perturbation (Forster *et al.*, 2007), the planet must increase its net energy loss to space in order to re-establish energy balance (with net energy loss being the difference between the outgoing long-wave (LW) radiation and net incoming short-wave (SW) radiation at the top-of-atmosphere (TOA)). Assuming that this increased energy loss is proportional to the surface temperature change ΔT, we can write

$$\Delta F = \lambda \Delta T + \Delta Q \tag{1}$$

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 new equilibrium state, there will be an imbalance, ΔQ, between radiative forcing and climate response. This imbalance represents the net heat flux into the system, with nearly all of this heat flux at present going into the ocean (Levitus *et al.*, 2005). Sustained forcing due to long-lived GHGs allows for significant exchange of heat to occur between the upper mixed layer and deep ocean, which delays the full surface temperature response by centuries-to-millennia. This delay is also a strong function of climate sensitivity (Hansen *et al.*,

[†]This refers to the time required for the feedbacks to become established, or 'felt' by the climate system in a significant way, following an imposed forcing.

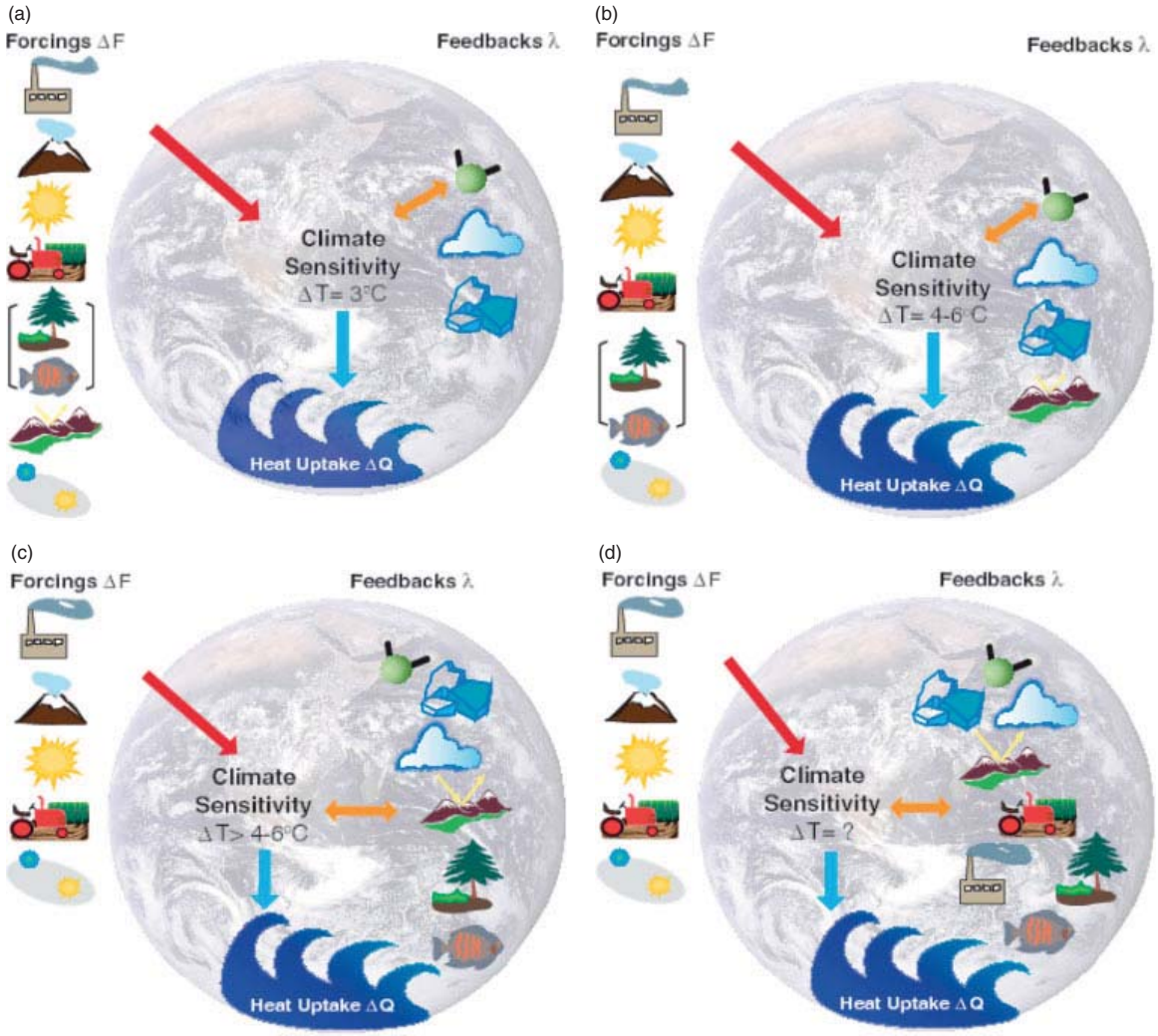

Figure 1. A climate forcing ΔF triggers a series of feedbacks (represented by the feedback parameter λ) which determine the resulting equilibrium global mean surface temperature change, or climate sensitivity, ΔT. Delay in this equilibrium temperature response due to ocean and cryosphere inertia leads to a net planetary heat uptake ΔQ. Different types of climate sensitivity are distinguished by the climate feedbacks that they include. (a) Fast Feedback Sensitivity: Climate sensitivity to an imposed external forcing depends solely on fast climate feedbacks due to changes in water vapour, 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, climate-related changes in terrestrial carbon sequestration, climate-related 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. The fast feedback sensitivity to a doubling of atmospheric CO_2 has been estimated to be about 3°C. (b) Earth System Sensitivity including Ice Sheet/Vegetation Albedo Feedbacks: If surface albedo changes from land ice and vegetation are regarded as a feedback, the climate sensitivity to a doubling of CO_2 increases to about 4–6°C. (c) Earth System Sensitivity additionally including Climate–GHG Feedbacks: If changes in atmospheric greenhouse gas (GHG) concentrations resulting from climate-related changes in terrestrial and ocean carbon sequestration are also regarded as a feedback, the $2\times CO_2$ climate sensitivity is higher still (>4–6°C). (d) Earth System Sensitivity additionally including Human Behaviour Feedbacks: In the most comprehensive type of climate sensitivity, changes in human activity (e.g. changes in fossil fuel burning, land use and land/ocean ecosystem management) in response to ongoing climate change are regarded as a feedback. (Note that human behaviour changes can be either a forcing or a feedback, since they can initiate Earth system change and also be a response to that change.)

1985). For short-lived forcings (e.g. volcanic aerosols), the deep ocean heat uptake is much smaller, and thus the full surface temperature response occurs much more rapidly as the upper mixed layer adjusts on a time-scale of months-to-years. At present, ΔQ (referred to herein as the ocean heat uptake) is estimated to be $0.58 \pm 0.15\,\mathrm{W\,m^{-2}}$ (Hansen *et al.*, 2011), 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 blackbody) 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 $\lambda_0 \approx 3.8\,\mathrm{W\,m^{-2}\,^{\circ}C^{-1}}$ (Appendix B). Therefore, *in the absence of any feedbacks* (i.e. $\lambda = \lambda_0$), a doubling of the atmospheric 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°C (Appendix B). 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 system.

3. Fast feedback sensitivity

The types of climate feedbacks that are operating (and their magnitudes) depend on the time-scale considered, the characteristics of the forcing (e.g. spatial pattern, spectral dependence), and the climate state when the forcing is applied. Fast feedbacks occurring on time-scales of decade(s) or less are associated with changes in atmospheric lapse rate, water vapour, clouds, sea ice, snow cover, and natural (i.e. non-anthropogenic) aerosols. One can then define the fast feedback climate sensitivity as the particular case in which only fast feedback processes act to modify the basic Planck response to a forcing (Figure 1(a)). The classic fast feedback sensitivity problem was defined by Charney (1979)[‡] who considered the response to a doubling of the atmospheric CO_2 concentration. It was concluded, based largely on a very limited number of 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 the fast feedback sensitivity based on the response to individual volcanic eruptions, climate change during the instrumental period (i.e. the last ∼150 years) and last millennium, Pleistocene glacial–interglacial transitions (e.g. from the last glacial maximum (LGM, ∼20 thousand years (ky) before present (BP)) to pre-industrial Holocene), and climate change occurring on longer time-scales such as the Cenozoic (the past 65.5 million years (My)) and even the Phanerozoic (the past 545 My). (Note that in these studies, any changes in land ice and vegetation were regarded as forcing; see Figure 1(a).) Combining evidence from this previous work suggests a most likely value and uncertainty range for the fast feedback sensitivity similar to those given by Charney, but with higher sensitivities difficult to rule out (Hegerl *et al.*, 2007; Knutti and Hegerl, 2008).

Uncertainty in the fast feedback sensitivity arises from several sources, and it is helpful to discuss these with the aid of Eq. (1). In empirical studies, the typical approach has been to calculate the climate feedback parameter λ using estimates of the radiative forcing ΔF and surface temperature change ΔT between two climate states. (We assume for the moment that two equilibrium states are considered, so that $\Delta Q = 0$.) The ratio $\Delta F_{2\times CO2}/\lambda$ then gives the climate sensitivity to a doubling of CO_2, where $\Delta F_{2\times CO2} = 3.7\,\mathrm{W\,m^{-2}}$ is the $2\times CO_2$ forcing. Such empirically derived sensitivity estimates are uncertain because past forcing and surface temperature change are uncertain. Additionally, it is assumed that λ inferred from past climate changes, which were driven by a variety of different forcings, can be used to reliably compute the climate sensitivity to a purely CO_2 forcing. This assumption is justified, however, provided that forcing 'efficacy' is appropriately accounted for (Hansen *et al.*, 2005)[§]. In GCM studies, model representation of individual feedback processes is the dominant source of uncertainty in the fast feedback sensitivity. In particular, cloud feedback has long contributed the most to this uncertainty (e.g. Charney, 1979; Hansen *et al.*, 1984; Soden and Held, 2006). One technique that has been employed to explore GCM uncertainty is the so-called perturbed physics ensemble (PPE: e.g. Murphy *et al.*, 2004; Sanderson *et al.*, 2008; Sanderson, 2011). In the PPE approach, poorly constrained model parameters (e.g. related to cloud processes) are varied over a plausible range of values, and resulting effects on the simulated climate feedbacks and sensitivity are assessed. Finally, the fast feedback sensitivity has been estimated from non-equilibrium states using both observations (e.g. from the instrumental period) and coupled atmosphere–ocean GCMs, in which case uncertainty in the ocean heat uptake ΔQ also comes into play. The term 'effective climate sensitivity' is often used to describe these estimates based on non-equilibrium conditions (Murphy, 1995).

4. Earth system sensitivity including ice sheet and vegetation albedo feedbacks

When calculating the fast feedback sensitivity, changes in continental ice sheets, and albedo effects of vegetation distribution/structure and the exposure of continental margins through changes in sea level, are either not considered or are included as part of the forcing. This is based on the long-standing notion that continental ice sheet changes occur so slowly (over several millennia) as to make them largely irrelevant to anthropogenic climate change. It was thus assumed that ice sheet/vegetation surface albedo feedbacks could be ignored. However, evidence from the palaeoclimatic record for sea-level changes of several metres per century (Thompson and Goldstein, 2005; Hearty *et al.*, 2007; Bard *et al.*, 2010; Miller *et al.*, 2011), as well as present-day observations of increasing melt and overall mass loss from Greenland and Antarctica (Rignot and Jacobs, 2002; Zwally *et al.*, 2002; Chen *et al.*, 2006; Tedesco, 2007; van den Broeke *et al.*, 2009; Wu *et al.*, 2010; Rignot, 2011; Zwally *et al.*, 2011), imply that ice sheet changes can occur more rapidly than previously recognized. Furthermore, both observation (proxy) based studies (e.g. Peteet *et al.*, 1994; Mann *et al.*, 2002; Bos *et al.*, 2005; Birks and Birks, 2008) and modelling studies (e.g. Jones *et al.*, 2009) indicate that significant vegetation response can occur on decadal-to-centennial time-scales. This suggests that ice sheet and vegetation albedo feedbacks should be included in the definition of climate sensitivity (Figure 1(b)). As noted above, this type of climate sensitivity is typically referred to as the Earth system sensitivity (e.g. Lunt *et al.*, 2010; Pagani *et al.*, 2010). An important point is that even though ice sheets are now believed to be capable of responding to climate warming more rapidly than previously thought, they are still quite lethargic, requiring centuries or longer to change their area significantly. However, this response is made possible by the very long lifetime of anthropogenic CO_2 (Archer *et al.*, 2009). Additionally, it is worth noting that ice sheet and vegetation changes may be important not just for their effect on surface albedo, but also because of other feedbacks they may induce such as changes in the ocean's thermohaline circulation (Swingedouw *et al.*, 2008; Goelzer *et al.*, 2011).

Hansen *et al.* (2008) estimated an Earth system sensitivity including ice sheet and vegetation albedo feedbacks of about 6°C for doubled CO_2. This is an average Earth system sensitivity for the range of climate states between glacial conditions and ice-free Earth, and thus it largely

[‡]Charney did not consider feedbacks associated with changes in natural aerosols. It is also interesting to note that he never actually used the term 'climate sensitivity'.

[§]The efficacy of a forcing agent is defined as the global surface temperature response to a unit forcing from that agent relative to the response to a unit forcing from CO_2.

reflects the changes that occurred during the Pleistocene glacial cycles. For smaller ice sheet changes, the sensitivity would be somewhat less. In this case, a useful palaeo-analogue for the future could be the mid-Pliocene warm period (∼3 My BP), which, relative to present-day, featured similar atmospheric CO_2 levels and considerably smaller changes in global ice volume compared to those which characterized the Pleistocene glacial cycles. Lunt *et al.* (2010) estimated an Earth system sensitivity for doubled CO_2 of 4–4.5°C for the mid-Pliocene relative to pre-industrial times. Although clearly smaller than the 6°C sensitivity given by Hansen *et al.* (2008), this nevertheless represents a significant enhancement (by ∼30–50%) of the fast feedback sensitivity.

Continued investigation is needed in order to better constrain the range of possible magnitudes and the time dependence of the ice sheet/vegetation feedback (e.g. through more careful reconstructions of glacial–interglacial ice sheet and vegetation changes). For instance, while the magnitude of atmospheric CO_2 changes was about the same between the last interglacial (∼125 ky 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 (e.g. Turney and Jones, 2010), indicating stronger amplifying feedbacks at work. This is supported by a larger sea-level change between the last interglacial and LGM (Kopp *et al.*, 2009), which suggests a stronger ice sheet feedback.

Further work is also needed in order to better constrain the time-scales over which the ice sheet feedback may become significant. Evidence for centennial time-scale ice sheet changes based on palaeo-sea-level records (see above) is largely derived from periods in Earth's history (e.g. the transition from the LGM to the Holocene) which featured greater amounts of ice than is available today (with much of this ice existing at relatively lower latitudes and elevations). One could argue that this would tend to favour a slower ice sheet response today than occurred during past warming. However, it must also be borne in mind that the current anthropogenic forcing greatly exceeds the forcing from orbital variations that drove past deglaciations, which might be expected to compensate to some extent (perhaps entirely) for the smaller present-day global ice volume. At the heart of the uncertainty surrounding ice sheet response time is the incomplete understanding of the dynamical processes (e.g. ice stream acceleration, ice shelf disintegration) that are thought to play a critical role in ice sheet decay (e.g. Dupont and Alley, 2006). Working to better understand these processes, and representing them in ice sheet models, are therefore crucial next steps toward narrowing the range of possible future ice sheet changes.

5. Adding climate–greenhouse-gas feedbacks to the Earth system sensitivity

In the Anthropocene, changes in atmospheric CO_2 concentrations are equivalent to anthropogenic emissions minus the net CO_2 uptake by the oceans and land surface (including the terrestrial biosphere). A portion of the ocean and land CO_2 uptake occurs rapidly following an anthropogenic CO_2 emission, as the added CO_2 equilibrates between the atmosphere and surface reservoirs. At present, about 57% of anthropogenic CO_2 emissions over the course of a year are taken up by the ocean and terrestrial biosphere (average annual uptake during the period 1959–2008: Le Quéré *et al.*, 2009). The remaining portion (43%) of annual anthropogenic CO_2 emissions that stays in the atmosphere is referred to as the 'airborne fraction'. The present-day oceans and land surface are therefore net sinks for anthropogenic CO_2. Because this CO_2 uptake occurs rapidly, it should be thought of as determining the magnitude of the forcing (i.e. rather than as a feedback). In other words, the rapid ocean and land CO_2 uptake fundamentally determines the airborne fraction of anthropogenic CO_2 emissions, and thus the atmospheric CO_2 concentrations (and associated TOA radiative imbalance) that are actually 'felt' by the system on climate change time-scales (i.e. decades and longer).

Thus, while present-day ocean and land CO_2 sinks are implicitly part of the forcing, any *changes in the magnitude of these sinks due to climate change* are a feedback (Figure 1(c)). Traditionally the latter were not considered or were also regarded as part of the forcing (Figure 1(a) and (b)). Both climate–carbon-cycle models (Friedlingstein *et al.*, 2006) and ice core records of atmospheric CO_2 concentrations during the Pleistocene (Lüthi *et al.*, 2008) indicate that the ability of the oceans and terrestrial biosphere to sequester CO_2 decreases with climate warming. This suggests that the present-day ocean and land CO_2 sinks will weaken in the coming decades as climate change progresses, signifying a positive climate–CO_2 feedback. We can also think of this feedback as an increase with time in the annual airborne fraction of anthropogenic CO_2 emissions (Friedlingstein *et al.*, 2006; Plattner *et al.*, 2008; Archer *et al.*, 2009). The strength of the climate–CO_2 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–CO_2 feedbacks (Friedlingstein *et al.*, 2006), which leads to an additional climate warming of between 0.1 and 1.5°C. In section 5.1, we will discuss in more detail the physical basis for the expected climate–CO_2 feedback.

On longer time-scales (ranging from thousands to hundreds of thousands of years), calcium carbonate ($CaCO_3$) neutralization and silicate weathering (both on the ocean floor and on land) will act to draw down atmospheric CO_2 concentrations (e.g. Archer *et al.*, 2009). While these processes are to some extent influenced by climate change (e.g. chemical weathering on land is enhanced under warmer and wetter conditions), they are fundamentally driven by the CO_2 increase itself. Additionally, it is clear that $CaCO_3$ neutralization and silicate weathering operate on very different time-scales than the characteristic time-scales of global mean surface temperature change. We can therefore think of these processes as acting to reduce the magnitude of anthropogenic CO_2 forcing over periods of many thousands of years. The implications of this for the Earth system sensitivity will be discussed in section 5.3.

5.1. *Climate–CO_2 feedbacks*

Present-day land and ocean carbon exchange with the atmosphere is expected to be affected both by the increase in atmospheric CO_2 concentration itself, and by the climate response 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 act to increase CO_2 sequestration rates. 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 (Reich *et al.*, 2006; Denman *et al.*, 2007; Hyvonen *et al.*, 2007; Gruber and Galloway, 2008; Heimann and Reichstein, 2008; Arneth *et al.*, 2010; Zaehle *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 wildfire and land-use change) determine the net carbon exchange 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; Kirschbaum, 2004; Knorr *et al.*, 2005; Davidson and Janssens, 2006). Other climate changes, in particular changes in the hydrological cycle (e.g. in drought frequency/severity), will also affect NPP and R_h, and thus it is critical to consider these changes as well. Additionally, warming-related increases in boreal forest fires (Soja *et al.*, 2007) and pests will likely offset at least a portion of the expected increase in NPP. Finally, it is important to consider how changes in anthropogenic land use and management may impact terrestrial ecosystem carbon exchange. 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 and energy 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 comprise the ocean's carbonate system. When CO_2 molecules are added to sea water, the net effect is a reaction with carbonate ion to form bicarbonate ion, which reduces the amount of carbonate molecules available to react with further CO_2 additions. This increases the pCO_2 of the sea water and thus decreases the ocean's 'buffering capacity' to draw down atmospheric CO_2 concentrations (Denman *et al.*, 2007). The buffering capacity is ultimately restored on multi-millennial time-scales by dissolution of $CaCO_3$ (Broecker and Takahashi, 1978; Ridgwell and Zeebe, 2005).

Atmospheric CO_2 uptake is also determined by the rate of the ocean's vertical mass 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 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 mass mixing as well as temperature and pH would also affect the biological component of the ocean's carbon cycle (Sarmiento *et al.*, 2004). This 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.

5.2. *Feedbacks between climate change and other greenhouse gases*

It is also important to consider how climate change may influence the sources and sinks of other GHGs besides CO_2 (e.g. Beerling *et al.*, 2011). 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). Increases in CH_4 during the industrial era produced the second-largest radiative forcing of the well-mixed GHGs 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 vapour (Hansen *et al.*, 2005). 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. Similarly, warming-induced permafrost thaw and thermokarst processes could increase landscape wetness and CH_4 emissions (Grosse *et al.*, 2011). However, undisturbed peatlands currently remove CO_2 from the atmosphere during photosynthesis and are hence a net sink for total carbon (including CO_2 and CH_4: Frolking *et al.*, 2011). Thus, any changes in CO_2 sequestration must also be factored in when determining the net carbon cycle feedback. It is generally expected that changes in CH_4 emissions may be important on decadal time-scales, but that on century-to-millennial time-scales CO_2 effects will dominate as a result of the much longer time required for atmospheric CO_2 concentrations to reach a new equilibrium following a perturbation to the peatland–atmosphere carbon exchange (Frolking and Roulet, 2007). 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. (Terrestrial clathrates likely constitute a much smaller pool of CH_4 than marine clathrates, and this CH_4 is less apt to be released into the atmosphere if/when clathrate destabilization occurs (Brook *et al.*, 2008).) While an abrupt release of methane as a result of marine clathrate destabilization appears very unlikely over the next century, there is likely to be an increase in the background rate of chronic CH_4 emission from clathrates during this time (Brook *et al.*, 2008).

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 (Erisman *et al.*, 2011). The direct impacts are associated with nitrous oxide (N_2O) and ozone (O_3). N_2O and O_3 are GHGs, and their increased abundance (in the troposphere for O_3) due to human activity has a warming effect. 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).

5.3. *Implications for Earth system sensitivity*

Since climate–GHG feedbacks are positive, the Earth system sensitivity is higher with these feedbacks included. Hansen and Sato (2012) find that including non-CO_2 GHG (CH_4 and N_2O) changes as a feedback increases the Earth system sensitivity to 8°C for doubled CO_2. They stress, though, that since this estimate is based on the LGM–Holocene transition, which featured a strong ice sheet/vegetation albedo feedback, it is likely on the high end of what is relevant in the Anthropocene. If (non-anthropogenic) atmospheric CO_2 responses to climate change are also counted as a feedback, the Earth system sensitivity would be higher still. At this point, however, it becomes difficult to define the sensitivity in a meaningful way based on the Pleistocene glacial cycles, since these cycles were driven by orbital variations which produced a negligible global mean forcing**.

Another consideration is whether the 'doubled CO_2' Earth system sensitivity is even relevant given that slow $CaCO_3$ neutralization and silicate weathering processes will draw down atmospheric CO_2 concentrations on multi-millennial time-scales. In other words, will a $2\times CO_2$ forcing be sustained for a long enough period of time to allow the full response of global surface temperature (as given by the $2\times CO_2$ sensitivity) to occur? Archer *et al.* (2009) find in their analysis of several carbon cycle models that for an instantaneous anthropogenic CO_2 release equivalent to 1000 Pg C (an amount that could be released by the end of the twenty-first century under business-as-usual emissions), atmospheric CO_2 concentrations fall below 560 ppm (i.e. double the pre-industrial concentration) within a few hundred years (see their Fig. 1). This is clearly shorter than the full response times of the oceans and ice sheets, indicating that in this case the $2\times CO_2$ Earth system sensitivity would likely overestimate the magnitude of future global warming. In this case, however, the CO_2 forcing remains greater than 1 W m^{-2} (CO_2 concentration greater than ~335 ppm) even after 10 ky (the length of the simulations analysed by Archer *et al.* (2009)). This suggests that it would be relevant for a 1000 Pg C emission to consider the Earth system sensitivity to a unit forcing in W m^{-2} (i.e. $1/\lambda$). For more extreme anthropogenic CO_2 emissions (5000 Pg C, corresponding approximately to the entire reservoir of fossil fuels), atmospheric CO_2 concentrations generally stay more than twice pre-industrial levels for 10 ky (Archer *et al.*, 2009), and thus the $2\times CO_2$ Earth system sensitivity would be an appropriate indicator of the magnitude of future warming.

Finally, it is important to point out that the above discussion of feedbacks involving the natural carbon and nitrogen cycles considers only feedbacks associated with changes in atmospheric GHGs. However, carbon and nitrogen cycle feedbacks also include changes in the atmospheric concentrations of natural aerosols (e.g. dust, carbonaceous particles from wildfires, NH_4 and NO_3 particles from ammonia chemistry) brought about by climate change. While natural aerosol feedbacks have the potential to be globally significant and tend to be negative, their net strength and even sign are highly variable and therefore uncertain (e.g. Carslaw *et al.*, 2010). We note, though, that these aerosol feedbacks are considered fast feedbacks (see section 3) and are hence included in the empirical estimates of Earth system sensitivity based on palaeodata that are given above.

6. Conclusions

Climate sensitivity is a concept that has evolved along with our understanding of the Earth system. This has resulted in several types of sensitivity that are distinguished by the Earth system feedbacks that they include, as discussed above and illustrated in Figure 1. In the traditional fast feedback sensitivity, the global mean surface temperature response to an externally imposed climate forcing is determined solely by fast climate feedbacks associated with changes in atmospheric lapse rate, water vapour, clouds, sea ice, snow cover, and natural aerosols. More comprehensive forms of the climate sensitivity including additional feedbacks are typically referred to as Earth system sensitivity. One type of Earth system sensitivity includes surface albedo feedbacks due to changes in continental ice sheets and vegetation, while a second type also incorporates climate–GHG feedbacks. We suggest that the latter is the most relevant form of climate sensitivity in the Anthropocene, since it includes all feedbacks that are expected to be important in determining the eventual (equilibrium) surface temperature response to the anthropogenic increase in GHG concentrations. The inclusion of climate–GHG feedbacks due to changes in the natural carbon sinks (Figure 1(c)) has the advantage of more directly linking anthropogenic GHG emissions with the ensuing global temperature increase, thus providing a truer indication of the climate sensitivity to human perturbations.

The pertinence of the Earth system sensitivity in the Anthropocene is further emphasized by Figure 2, which shows the climate sensitivity to a doubling of the CO_2 concentration versus the time required to achieve this equilibrium temperature response. If only fast climate feedbacks are considered (blue circle in Figure 2), the $2\times CO_2$ sensitivity is about 3°C, as has been inferred from both climate models and observations of climate change during the instrumental period (see section 3). One notes from Figure 2, however, that this 3°C warming would take several centuries to about a millennium to be realized (e.g. Hansen *et al.*, 2011), due to exchange of heat between the mixed layer and deep ocean. This slow response time of the Earth system to long-lived forcings such as anthropogenic CO_2, which is a consequence of the large heat capacity of the deep ocean, enables additional feedbacks associated with changes in the natural carbon cycle, continental ice sheets and vegetation to come into play (as indicated by dashed lines in Figure 2). While these (positive) feedbacks are not operating in atmosphere–ocean GCMs, they are operating in the real world, implying that the real-world climate sensitivity to doubled CO_2 is higher than in models. The relevant form of the $2\times CO_2$ sensitivity in the real world is, therefore, the Earth system sensitivity (green and tan circles in Figure 2).

Climate sensitivity in the Anthropocene is therefore higher than the fast feedback sensitivity that has typically

**From Eq. (1), this implies an extremely small value of λ, and thus an extremely large value of $\Delta F_{2\times CO_2}/\lambda$, the $2\times CO_2$ sensitivity.

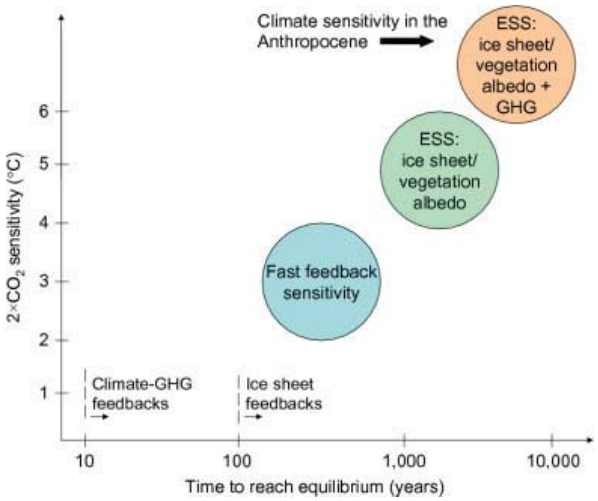

Figure 2. Schematic showing the climate sensitivity (°C) to an instantaneous doubling of the atmospheric CO_2 concentration versus the time required to achieve this equilibrium surface temperature response (in years since CO_2 doubling). Different coloured circles represent the three main types of climate sensitivity discussed in the text, specifically the fast feedback sensitivity, the Earth system sensitivity (ESS) including ice sheet/vegetation albedo feedbacks, and the ESS additionally including climate-greenhouse gas (GHG) feedbacks. Dashed lines indicate the approximate time-scales on which climate–GHG feedbacks and ice-sheet albedo feedbacks are expected to become significant (decades or longer and centuries or longer, respectively). We suggest that the ESS including both ice sheet/vegetation albedo and climate–GHG feedbacks is the most relevant form of climate sensitivity in the Anthropocene.

been assumed. A similar conclusion has been reached by several other authors (e.g. Lashof, 1989; Hansen *et al.*, 2008; Lunt *et al.*, 2010; Pagani *et al.*, 2010; Kiehl, 2011; Park and Royer, 2011), all of whom pointed to the relevance of the Earth system sensitivity for future anthropogenic climate change. The higher Earth system sensitivity implies an even longer response time of at least several millennia (Figure 2), since the time needed to reach a new equilibrium state following a CO_2 doubling increases nonlinearly with climate sensitivity (Hansen *et al.*, 1985). There are also implications for the amount of committed warming that is still 'in the pipeline' as a result of past forcing not yet responded to[††]. It is commonly believed that global warming should be kept below 2°C relative to pre-industrial times, as greater amounts of warming could lead to dangerous climate change impacts (e.g. European Council, 2005). Given that ∼0.8°C of warming has already occurred (Hansen *et al.*, 2010), this would require that the committed warming not exceed 1.2°C. For the present-day planetary energy imbalance of 0.58 W m^{-2}, this means that the climate sensitivity for doubled CO_2 would need to be less than about 7°C ($\lambda > 0.5\,\mathrm{W\,m^{-2}\,^{\circ}C^{-1}}$). While the fast feedback sensitivity is very likely less than 7°C, it is not clear that this is the case for the Earth system sensitivity including ice sheet and climate–GHG feedbacks. The higher Earth system sensitivity thus implies a real possibility of exceeding the 2°C global warming threshold if atmospheric GHG concentrations are sustained at or above present-day levels[‡‡]. This needs to be communicated clearly to policymakers and to the general public in order to ensure appropriately informed decisions about future GHG stabilization.

7. Future directions

In the introduction to this article, we defined climate sensitivity in the most general way as the global mean surface temperature response to an externally imposed climate forcing. In subsequent discussions, we then typically referred to the sensitivity to the canonical forcing associated with doubling the atmospheric CO_2 concentration. While the doubled CO_2 forcing may indeed remain the benchmark forcing, it is important to point out that defining climate sensitivity (i.e. Earth system sensitivity) to explicitly include climate–CO_2 feedbacks has implications for how we interpret and actually calculate the $2\times CO_2$ sensitivity. In particular, in the Anthropocene when atmospheric CO_2 concentration changes are both a forcing and a feedback, it is necessary to separate the total CO_2 change along these lines when calculating the climate sensitivity (from empirical data or Earth system model (ESM) output). The forcing then becomes the atmospheric CO_2 concentration change that would result directly from the anthropogenic emissions in the absence of any climate-related changes in the natural carbon sinks (i.e. constant airborne fraction), while the feedback is the difference between this and the total (actual) atmospheric CO_2 change (i.e. CO_2 change due to change in airborne fraction). This separation of the total CO_2 change into climate forcing and climate–CO_2 feedback (assumed to be positive) thus implies a greater climate sensitivity. However, it is important to bear in mind that on very long time-scales, the forcing will be reduced as a result of $CaCO_3$ neutralization and silicate weathering (see section 5). This casts some doubt on the utility of the $2\times CO_2$ sensitivity for anything other than very large emissions of anthropogenic CO_2.

Estimating Earth system sensitivity including ice sheet/vegetation albedo and climate–GHG feedbacks, in practice, is a significant challenge. As noted in section 5, this is difficult to do using Pleistocene glacial–interglacial transitions because the negligible global mean forcing from orbital variations makes the problem ill-posed. One might alternatively look to the instrumental period to estimate the Earth system sensitivity. The problem with this approach, however, is that ice sheet and climate–GHG feedbacks have yet to become globally significant, as evidenced by the relatively small changes that have been observed in Greenland and Antarctic ice sheet area and in the annual airborne fraction of anthropogenic CO_2 emissions. (There are of course other problems with the instrumental period, such as uncertainty in the anthropogenic aerosol forcing and ocean heat uptake.) Our ability to estimate Earth system sensitivity from models is also limited. ESMs include an interactive carbon cycle, and thus are capable of simulating climate–CO_2 feedbacks. However, the representation of the carbon cycle in these models does not account for certain processes that are

[††]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 change), but to the response of ecosystems as well (Jones *et al.*, 2009).

[‡‡]That current GHG levels may be problematic is further supported by the recent finding that global sea level during the mid-Pliocene, a time with atmospheric CO_2 levels similar to today, was about 25 m higher than at present (Rohling *et al.*, 2009).

likely to be important, such as interactions with the nitrogen cycle. Additionally, current ESMs are only beginning to incorporate interactive continental ice sheets, and do not yet represent the ice sheet dynamical processes that are expected to be critical for the ice sheet albedo feedback. Quantifying Earth system sensitivity including all relevant feedbacks therefore remains a high priority for future research.

Finally, it is interesting to speculate whether evolving knowledge of the Earth system will eventually suggest that still other feedbacks should be included in the definition of climate sensitivity. One particularly intriguing possibility is that human activity and its changes through time (e.g. changes in fossil fuel burning, land use and land/ocean ecosystem management) could be regarded as a feedback (Figure 1(d)). In this framework, humans would respond to anthropogenically forced climate change by altering their behaviour (e.g. Lashof, 1989), thus producing feedbacks that affect the natural system. (Traditionally, this response would instead be seen as a change in forcing.) This is already becoming a reality in models as GCMs are being coupled to integrated assessment models that represent various aspects of human activity such as energy use and land use (e.g. Prinn, 2012). The concept of an anthropogenic feedback requires us to think about climate sensitivity in a very different way. We conclude, though, that whether viewed as a forcing or a feedback, future changes in human activity will remain the single greatest source of uncertainty in climate change projections.

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Appendix A. Definition of radiative forcing

Various radiative forcing definitions have been adopted in the scientific literature. The simplest of these is the instantaneous forcing, which is defined as the radiative flux change at the tropopause after the forcing agent has been introduced with the climate held fixed. Another forcing definition, and the one traditionally adopted by the Intergovernmental Panel on Climate Change (IPCC), is the adjusted forcing, which is the flux change at the top-of-atmosphere (TOA) and throughout the stratosphere after stratospheric temperatures have been allowed to adjust radiatively to the presence of the forcing agent. Alternative methods of calculating the forcing further allow for adjustment of tropospheric and land surface temperatures, and for various carbon dioxide (CO_2) and aerosol effects on clouds. See Liepert (2010) for a recent review of this topic.

Appendix B. Calculating the Planck response of the Earth's long-wave emission

The TOA radiative balance can be written as $S = \sigma T_e^4$, where $S = 239\,\mathrm{W\,m^{-2}}$ is the solar radiation absorbed by Earth and σT_e^4 is the outgoing long-wave (LW) radiation, with $\sigma = 5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$ being the Stefan–Boltzmann constant. This relationship allows one to calculate the effective emission temperature of the Earth as $T_e \approx 255\,\mathrm{K}$. T_e is also the physical temperature at some mean level of emission to space, which, in the current atmosphere, occurs at an altitude of about 6 km (Hansen *et al.*, 1984). Following a positive radiative forcing, the outgoing LW radiation must increase in order to restore the TOA energy balance. This Planck response of the LW emission is obtained simply by differentiating the emission with respect to T_e: $\lambda_0 = \mathrm{d}(\sigma T_e^4)/\mathrm{d}T_e = 4\sigma T_e^3 \approx 3.8\,\mathrm{W\,m^{-2}\,{}^\circ C^{-1}}$. Therefore, a doubling of the atmospheric CO_2 concentration, which represents a forcing $\Delta F = 3.7\,\mathrm{W\,m^{-2}}$ (Forster *et al.*, 2007), would require T_e to increase by $\Delta F/\lambda_0 \approx 1^\circ\mathrm{C}$. If we assume no change in the atmospheric lapse rate, this would also be the magnitude of temperature increase at the surface. In other words, in the absence of any feedbacks (i.e. considering only the Planck response of the Earth's LW emission), the climate sensitivity to a doubling of CO_2 would be about 1°C.

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