Climate Sensitivity in the Anthropocene

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Supplementary information

1. Definition of radiative forcing and importance of non-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 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.

The traditional paradigm for climate sensitivity tends to focus exclusively on *radiative* forcings, without consideration of how anthropogenic (or natural) perturbations may affect the non-radiative energy fluxes in the surface energy budget that also determine the surface temperature. An illustrative example of the importance of these non-radiative (latent and sensible heat) fluxes is provided by considering the effects of doubling atmospheric CO_2. In response to a CO_2 doubling, the radiative heating of the Earth system (i.e., the TOA forcing) increases by 3.3 W m^{-2} (Gregory and Webb, 2008), with most of this additional heating (2.4 W m^{-2}) occurring within the troposphere and the remaining portion (0.9 W m^{-2}) occurring at the surface (Andrews et al., 2009). Since the heat capacity of the troposphere is small and can effectively be taken as zero, the gain of radiative energy due to increasing CO_2 must be compensated for by an equivalent loss of non-radiative energy, which occurs via a reduction of the upward latent and sensible heat fluxes from the surface. Thus, from a surface energy budget perspective, the non-radiative component of CO_2 forcing exceeds the radiative component of the forcing (2.4 and 0.9 W m^{-2}, respectively). Other climate forcings (e.g., anthropogenic aerosol changes; Ming and Ramaswamy, 2009; Ming et al., 2010) and feedbacks (Previdi and Liepert, 2011) similarly induce a significant alteration of the non-radiative energy exchange between the surface and atmosphere (see also Previdi, 2010). The bulk of this alteration occurs in the surface latent heat flux (Andrews et al., 2009; Previdi and Liepert, 2011), indicating that the global water cycle and its changes are fundamental to understanding climate sensitivity. This point is further emphasized by the fact that most feedbacks (e.g., due to changes in water vapor, clouds, sea ice and land ice) are directly associated with water and changes in its phase and storage in different components of the climate system.

2. Calculating the Planck response of the Earth's longwave emission

The TOA radiative balance can be written as $S = \sigma T_e^4$, where S = 239 W m^{-2} is the solar radiation absorbed by Earth and σT_e^4 is the outgoing longwave (LW) radiation, with σ = 5.67×10^{-8} W m^{-2} K^{-4} being the Stefan-Boltzmann constant. This relationship allows one to calculate the effective emission temperature of the Earth as $T_e \approx 255$ 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: $\lambda_0 = d(\sigma T_e^4)/dT_e = 4\sigma T_e^3$.

3. Pleistocene carbon cycle changes

Paleodata show greenhouse gas (GHG) changes lagging temperature changes by several hundred years (Caillon et al., 2003; Mudelsee, 2001), thus indicating that the former are a feedback on climate change. If GHG changes are regarded as a feedback, then the large amplitude of Pleistocene glacial-interglacial temperature changes would imply an extremely high climate sensitivity, since the global mean forcing due to orbital variations (which are the ultimate driver of the glacial cycles; Hays et al., 1976) is a negligible fraction of 1 W m^{-2} (Hansen et al., 2008).

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