GEOPHYSICAL RESEARCH LETTERS, VOL. ???, XXXX, DOI:10.1029/,

Climate commitment in an uncertain world

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Climate commitment—the warming that would still occur given no further human influence—is a fundamental metric for both science and policy. It informs us of the minimum climate change we face and, moreover, depends only on our knowledge of the natural climate system. Studies of the climate commitment due to CO_2 find that global temperature would remain near current levels, or even decrease slightly, in the millennium following the cessation of emissions. However, this result overlooks the important role of the non-CO_2 greenhouse gases and aerosols. This paper shows that global energetics require an immediate and significant warming following the cessation of emissions as aerosols are quickly washed from the atmosphere, and the large uncertainty in current aerosol radiative forcing implies a large uncertainty in the climate commitment. Fundamental constraints preclude Earth returning to pre-industrial temperatures for the indefinite future. These same constraints mean that observations are currently unable to eliminate the possibility that we are already beyond the point where the ultimate warming will exceed dangerous levels. Models produce a narrower range of climate commitment, but undersample observed forcing constraints.

1. Introduction

Our ability to predict future climate changes rests fundamentally on two factors: firstly, how our future human activities will influence climate forcing and secondly, how our models of the climate system translate that forcing into climate change. The first factor depends on societal choices beyond the scope of science. The second factor depends on our confidence in the climate models. In turn, this confidence is predicated on the ability of the models to reproduce past climate changes, given our knowledge of previous human (and other) influences.

The concept of a 'climate commitment'—the climate change that would still occur given no further human influence—has proven useful in distinguishing between these two factors of climate prediction. It allows for a clear separation between the uncertainties in our physical climate models, which we wish to study, and the highly-uncertain future human influence on climate. The climate commitment can also be regarded as the minimum climate change we are consigned to because of human activities already undertaken.

Early efforts to estimate climate commitment considered the additional warming that occurs as the climate system comes into equilibrium with the present atmospheric composition and radiative forcing. Under this assumption, an additional warming of about 0.6°C is 'in the pipeline' due to the thermal inertia of the world oceans [*Wigley*, 2005; *Meehl et al.*, 2005; *Hansen et al.*, 2005], committing us to future climate change that approaches 'dangerous' levels [*Ramanathan and Feng*, 2008].

There has recently been a resurgence of interest in the climate commitment [*Ramanathan and Feng*, 2008; *Hare and Meinshausen*, 2006; *Plattner et al.*, 2008; *Solomon* *et al.*, 2009; *Matthews and Weaver*, 2010] in which an alternative, 'zero emissions', definition has been proposed. Under zero emissions, the atmospheric composition changes according to natural processes, and future warming is determined by only the physical inertia of the climate system and the residual greenhouse gas climate forcing. *Matthews and Weaver* [2010] argue that this definition is the correct measure of the present climate commitment. They make the worthwhile and important point that the previous measure—constant climate forcing—conflates the physical response of the climate system to past emissions with the response to the future emissions that are necessary to maintain a constant atmospheric composition.

Several studies [*Plattner et al.*, 2008; *Solomon et al.*, 2009; *Matthews and Weaver*, 2010] consider the zero emissions commitment with respect to CO_2. Carbon dioxide is naturally removed from the atmosphere on multiple time scales. Under zero emissions, CO_2 would fall off to about 40% of its peak enhancement above pre-industrial levels within a few centuries [*Solomon et al.*, 2009], while full recovery would occur over hundreds of thousands of years [*Archer*, 2005]. Effectively then, this residual 40% defines the ultimate radiative forcing ($\equiv R_\infty$) with which the climate must come into equilibrium. In such a zero emissions scenario, global average surface temperature is projected to remain near current levels, or even decrease slightly, in the millennium following the cessation of emissions [*Plattner et al.*, 2008; *Solomon et al.*, 2009; *Matthews and Weaver*, 2010]. However, these studies have overlooked the important role of the non-CO_2 greenhouse gases (such as methane and nitrous oxide) and aerosols. Aerosols are widely known to be one of the chief uncertainties in the modern climate, and make a considerable difference to the answer. *Ramanathan and Feng* [2008] do consider the effect of removing anthropogenic aerosols, however they fix CO_2 at modern levels. The full consequences of the cessation of human activities must include both influences.

2. Transient and ultimate climate commitment

Following the elimination of emissions, aerosols would fall to their pre-industrial levels on time scales of days to weeks [*Forster et al.*, 2007], while the non-CO_2 greenhouse gases would persist for decades to centuries [*Forster et al.*, 2007; *Solomon et al.*, 2009]. The sudden loss of the cooling effect of aerosols would result in a rapid transient warming as the surface temperature adjusts to the full greenhouse gas radiative forcing. Due to this significant transient warming, we propose two separate measures of climate commitment: a 'transient commitment', defined by the peak temperature following the cessation of emissions; and an 'ultimate commitment', defined by the temperature once the climate system has fully equilibrated with the persistent fraction of the CO_2 radiative forcing.

How well constrained is the climate commitment? Conservation of energy must obviously apply to the global energy budget, a linearization of which is

$$H = R - \lambda^{-1} T, \tag{1}$$

where λ is the climate sensitivity parameter, T is the global average surface temperature (above pre-industrial), R is the radiative forcing, and H is the ocean heat uptake.

For a permanent forcing R_∞, H must ultimately go to zero giving an ultimate commitment of

$$T_\infty = \lambda R_\infty. \tag{2}$$

Eliminating λ gives

$$T_\infty = \left(\frac{R_\infty}{R-H}\right)T. \tag{3}$$

Thus, T_∞ depends only on observed constraints (T, H, and R) and the ultimate forcing (R_∞). For the current climate, T is $0.76 \pm 0.11\,°\mathrm{C}$ (1σ) [*Trenberth et al.*, 2007] and H is 0.74 ± 0.08 W m^{-2} (1σ) [*Lyman et al.*, 2010; *Purkey and Johnson*, 2010].

The Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) outlines constraints on R. Total anthropogenic radiative forcing is approximately 1.6 W m^{-2}, with a 90% confidence range of 0.6 W m^{-2} to 2.4 W m^{-2} [*Forster et al.*, 2007]. Therefore, T and H are well constrained, R less so. Uncertainties in aerosols (-0.5 W m^{-2} to -2.2 W m^{-2}) [*Forster et al.*, 2007] dominate the uncertainty in R, and thus dominate the uncertainty in T_∞.

The time evolution of climate requires a representation of the ocean, for which we use a simple upwelling-diffusion model. The model is the same as that in *Baker and Roe* [2009], which is similar in form to those used in previous studies (e.g., *Hoffert et al.* [1980]; *Raper et al.* [2001]). All parameters are as described in *Baker and Roe* [2009], except R and λ, which we vary as described below. Such models are robust, and successfully reproduce observations of ocean heat uptake at the global scale [*Raper et al.*, 2001].

3. Results

The weak bounds on aerosols means a broad envelope of uncertainty in total forcing over the industrial era. This is illustrated in Fig. 1a, where an idealized representation of forcing trends has been employed. Forcing reaches its modern value in year 200, and from then on a climate commitment scenario is assumed. Once emissions are terminated,

R is governed by the respective decay time scales of the various atmospheric constituents (see Fig. 1).

From Eq. 1 the relatively strong constraints on T and H mean that R and λ can be thought of as pairs wherein strong (weak) aerosol forcing is balanced by high (low) climate sensitivity. This compensation occurs within AR4 and older models [*Schwartz et al.*, 2007; *Kiehl*, 2007; *Knutti*, 2008]. Figure 1b shows temperature trajectories for pairs of R and λ, whereby past temperature trends are approximately reproduced. It is a graphical representation of the inherent trade-off between uncertainties in climate forcing and uncertainties in global temperature following the cessation of emissions: even though past temperature changes are well constrained and future forcing (under zero emissions) well understood, uncertainty in past forcing implies uncertainty in future temperatures.

It is important to emphasize that R and λ are not independent. In other words, a high climate sensitivity and a low aerosol forcing are inconsistent with the observed constraints on surface temperature and ocean heat uptake. Two recent studies that consider the effects of the loss of aerosols [*Hare and Meinshausen*, 2006; *Ramanathan and Feng*, 2008] treat R and λ as independent, and also fail to span the full range of either R or λ. This has the effect of producing a narrower range of climate commitment than allowed by propagating the observed constraints through Eq. 3.

We next reproduce and explain the results of previous studies [*Plattner et al.*, 2008; *Solomon et al.*, 2009; *Matthews and Weaver*, 2010] that considered climate commitment with respect to only CO_2 emissions (non-CO_2 greenhouse gases and aerosols remain at their modern concentrations). For modal estimates of modern radiative forcing, this gives

$R_\infty \approx 0.8$ W m^{-2}. The dashed black lines in Fig. 1 show this forcing and the response—a gentle decline in temperature following the cessation of CO_2 emissions. The result follows directly from surface energetics (Eq. 3): R_∞ is very near the modal value of the current surface forcing ($R - H \approx 1$ W m^{-2}) so the ratio of forcings (i.e., $R_\infty/(R - H)$) and therefore the ratio of the responses (i.e., T_∞/T) is near, but slightly less than, one.

Turning now to the case in which all anthropogenic emissions cease, there is an immediate unmasking of greenhouse gas forcing as aerosols are quickly washed from the atmosphere. The effect is an abrupt rise in climate forcing (Fig. 1a) to a peak value of around 2.7 W m^{-2}, which is relatively well constrained as it depends only on greenhouse gases. The response is a rapid warming (Fig. 1b), with a transient commitment of up to 0.9 °C above the modern temperature. Thereafter, forcing declines over the next few centuries as greenhouse gases are partially, but not completely, removed from the atmosphere. At the low end of the climate response, temperature falls to less than half of its peak value. At the high end, temperature continues to increase because the system has not yet attained equilibrium due to the long adjustment time scales of high sensitivity systems [*Baker and Roe*, 2009].

We note that while simple upwelling-diffusive climate models, such as the one used here, are able to reproduced observed climate trends, they do not accurately capture the complexities of ocean heat uptake at the regional scale [*Gregory*, 2000] and likely underestimate the long-term temperature response to forcing [*Winton et al.*, 2009]. While the details of any particular temperature trajectory are model dependent, the overall form of the temperature response is a fundamental consequence of three basic and robust climate properties: the unmasking of climate forcing by the loss of anthropogenic aerosols, the long lifetime of greenhouse gases in the atmosphere, and the thermal inertia of the ocean.

The long-term temperature response depends only on modern surface energetics and R_∞. Figure 1 accounts only for uncertainties in aerosols. This gives an ultimate commitment (above pre-industrial) of $T_\infty = 0.6\,^\circ$C with a 90% confidence range of $0.3\,^\circ$C to $7.2\,^\circ$C, which follows directly from Eq. 3 or by integrating the climate model to equilibrium.

The lower bound on climate commitment is robust due to the form of Eq. 3. On the other hand, the upper bound is very sensitive to uncertainties in observed global energetics and R_∞. We do not account here for uncertainties in the biogeochemical cycle (e.g., uncertainty in the lifetimes of greenhouse gases or the residual atmospheric CO_2 concentration). Moreover, following the IPCC, we have taken the 90% confidence interval on aerosol climate forcing: if one were to factor in other sources of uncertainty, in either ocean heat uptake or greenhouse gas forcing, or use more conventional statistical bounds (i.e., a 95% range), one could not rule out the disconcerting possibility that the observed 20th century warming has transpired with little to no effective surface forcing (i.e., $R \approx H$).

4. Discussion

The above analysis showed that current observational constraints allow the possibility of a very large climate commitment. Do narrower bounds exist? The ultimate commitment can alternatively be expressed as a function of λ (Eq. 2), reasonable bounds on which can be inferred from IPCC AR4 in terms of a 'likely' (> 66% probability) and 'very likely' ($>90\%$ probability) range for climate sensitivity [*Hegerl et al.*, 2007]. Exploiting the fundamental relationship between R and λ, and reversing the above arguments, these IPCC constraints on λ provide constraints on R (Fig. 2a). Any value of λ within the IPCC range still implies a significant transient warming (Fig. 2b), and there remains a substantial uncertainty in the ultimate commitment (though the range is smaller than that based on observational constraints).

The ability of the IPCC AR4 fully coupled climate models (hereafter AR4 models) to reproduce 20th century surface temperature [*Knutti*, 2008] and ocean heat uptake [*Plattner et al.*, 2008], under substantial aerosol uncertainty, has been suggested to give a false sense of the accuracy with which future climate can be predicted [*Schwartz et al.*, 2007]. However, AR4 models have achieved consistency with the observational record, in part, through compensation between R and λ [*Knutti*, 2008]. As argued by *Knutti* [2008], such model tuning—whether explicit or implicit—is not problematic provided that we interpret models as conditional on observations. In other words, models satisfy Eq. 1 subject to relatively tight constraints on T and H. Accurate simulation of 20th century climate may then be viewed as a necessary, but not sufficient, condition for the ability to simulate future climate, and does not alone create overconfidence in model skill. Indeed, the light blue trajectories in Fig. 2b clearly demonstrate the ability, with a model, to reproduce the 20th century temperature record yet still span the full range of uncertainty in climate commitment as allowed by observations.

The difference between the AR4 model range of climate commitment and the range allowed by observations can instead be attributed to an inconsistency between R in models and R in observations—the range of forcing among the different AR4 models [*Knutti*, 2008] spans only the 'likely' range of forcing in Fig. 2a. How can models and observations be reconciled? One way would be to achieve substantially more accurate observations of the Earth's radiative budget. In particular, emphasis should be placed on ruling out the very low values of R that correspond to very high values of committed warming.

The alternative approach is to create populations of climate models that deliberately exploit tuning to fully span the uncertainty in climate forcing (and the implied range of climate sensitivity necessary to reproduce the observed temperature record), and then to demonstrate that some pairs of R and λ are inconsistent with some aspect of either the instrumental record (e.g., interannual variability, seasonal variability, spatial patterns of warming, or volcanic eruptions), or reconstructions of past climates (see *Hegerl et al.* [2007]; *Knutti and Hegerl* [2008]; *Edwards et al.* [2007] and references therein). Studies that pursue this approach produce a variety of distributions for climate sensitivity, many narrower than that inferred from observational constraints, some narrower than even the IPCC 'likely' range [*Allen et al.*, 2007; *Hegerl et al.*, 2007; *Edwards et al.*, 2007; *Knutti and Hegerl*, 2008]. An implication then, would be that the range of uncertainty in the climate commitment could be narrowed as well. Achieving convergence among these different distributions depends on understanding the differing assumptions and structural uncertainties in, and the interdependence of, the respective frameworks [*Frame et al.*, 2005; *Allen et al.*, 2007; *Knutti and Hegerl*, 2008; *Knutti*, 2010]. Arguably, an important measure of the value added by models will be when the consensus is reached that such studies provide narrower constraints on the modern climate forcing than that currently provided by direct observations. The discrepancy between the reported ranges of uncertainty in climate sensitivity and observations of aerosol forcing is an important one for future rounds of the IPCC reports to resolve. Until then, model-based estimates should be treated carefully, and probably represent an undersampling of the possible climate commitment.

5. Conclusions

The results presented here depend only on three straightforward and well-understood aspects of climate: the net cooling effect of aerosols, the large spread of uncertainty in aerosol forcing (or, equivalently, climate sensitivity), and the long atmospheric lifetimes CO_2 and other greenhouse gases. In combination they lead to considerable uncertainty in the transient and ultimate climate commitments.

Our focus on the present climate commitment leads to one particular value of R_∞. Of course, in any practical scenario, emissions will continue and R_∞ will grow. In turn, the transient and ultimate climate commitments will increase and become more uncertain. Inasmuch as a substantially improved understanding of the role of aerosols in climate remains elusive, so will our ability to constrain future climate. In order to rule out the possibility that we already face a disturbingly large climate commitment, we need to rule out the possibility that the observed climate change has been driven by a climate forcing at the lower end of the range that is currently permitted by observations.

Acknowledgments. We thank M. B. Baker, D. Battisti, C. M. Bitz, E. Steig, and L. Thompson for insightful discussions; two anonymous reviewers for constructive suggestions that greatly improved the manuscript; and Noah Diffenbaugh, the editor.

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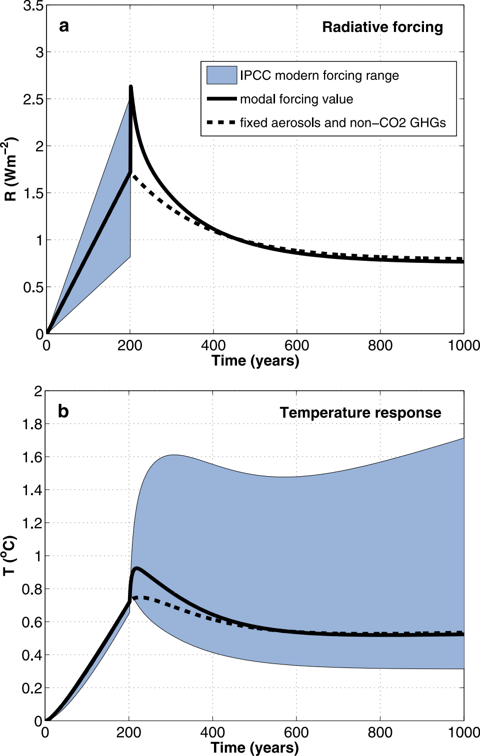
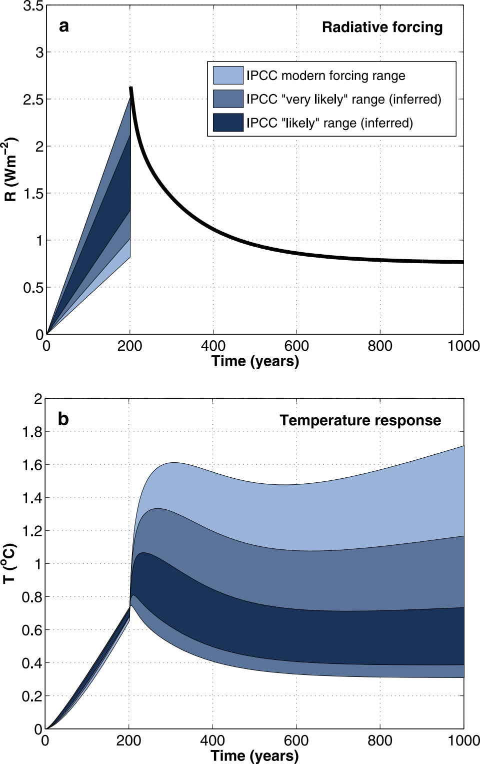
Figure 1. Observational constraints on climate forcing and temperature response. **a**, Idealized representation of forcing trends. Forcing is ramped linearly to its modern value in year 200, and a zero emissions scenario is assumed thereafter. Upon zero emissions, aerosols and tropospheric ozone are specified to fall to pre-industrial levels immediately. Long-lived greenhouse gases decline at their respective (e-folding) time scales [*Forster et al.*, 2007]: 12 years for methane; 114 years for nitrous oxide; 75 years (a representative lifetime) for halocarbons. Carbon dioxide falls to 40% of its peak value (above pre-industrial) with a decay time scale of 170 years [*Forster et al.*, 2007]. Radiative forcing is calculated using the simplified expressions of *Myhre et al.* [1998]. The light blue shading is the 90% confidence interval on trajectories of R as allowed by observations, where only uncertainty in aerosols is considered. The solid black line shows the modal value of R. The dashed black line shows a scenario in which aerosols and non-CO_2 greenhouse gases are held fixed at their modern concentrations upon the elimination of CO_2 emissions. **b**, As for **a**, but modeled temperature response. Values of λ have been paired with values of R so that individual temperature trajectories are tightly constrained, analogous to the situation for modern observations.

Figure 2. Illustration of what IPCC bounds on climate sensitivity imply for constraints on past climate forcing and future temperature response. **a**, Radiative forcing and **b**, Temperature response, as in Fig. 1. The dark blue shading shows the IPCC AR4 'likely' range of climate sensitivity (2 °C to 4.5 °C). The medium blue shading shows the IPCC AR4 'very likely' range of climate sensitivity (1.5 °C to 10 °C—IPCC AR4 [*Hegerl et al.*, 2007] truncates the probability distributions of climate sensitivity at 10 °C so we take this value as representative of the upper bound on the 'very likely' range). For comparison, the light blue shading shows the 90% confidence interval as allowed by observations, as in Fig. 1. A wedge in the lower range of possible forcing translates to a wedge in the higher range of possible temperature response.



a
Radiative forcing
IPCC modern forcing range
modal forcing value
fixed aerosols and non−CO2 GHGs
R (Wm^{-2})
Time (years)
b
Temperature response
T (^{o}C)
Time (years)



a
Radiative forcing
IPCC modern forcing range
IPCC "very likely" range (inferred)
IPCC "likely" range (inferred)
R (Wm^{-2})
3.5
3
2.5
2
1.5
1
0.5
0
0
200
400
600
800
1000
Time (years)
b
Temperature response
T (°C)
2
1.8
1.6
1.4
1.2
1
0.8
0.6
0.4
0.2
0
0
200
400
600
800
1000
Time (years)