# Why Hasn't Earth Warmed as Much as Expected?

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ABSTRACT

The observed increase in global mean surface temperature (GMST) over the industrial era is less than 40% of that expected from observed increases in long-lived greenhouse gases together with the best-estimate equilibrium climate sensitivity given by the 2007 Assessment Report of the Intergovernmental Panel on Climate Change (IPCC). Possible reasons for this warming discrepancy are systematically examined here. The warming discrepancy is found to be due mainly to some combination of two factors: the IPCC best estimate of climate sensitivity being too high and/or the greenhouse gas forcing being partially offset by forcing by increased concentrations of atmospheric aerosols; the increase in global heat content due to thermal disequilibrium accounts for less than 25% of the discrepancy, and cooling by natural temperature variation can account for only about 15%. Current uncertainty in climate sensitivity is shown to preclude determining the amount of future fossil fuel $CO_2$ emissions that would be compatible with any chosen maximum allowable increase in GMST; even the sign of such allowable future emissions is unconstrained. Resolving this situation, by empirical determination of the earth's climate sensitivity from the historical record over the industrial period or through use of climate models whose accuracy is evaluated by their performance over this period, is shown to require substantial reduction in the uncertainty of aerosol forcing over this period.

## 1. Introduction

Warming of the earth's climate system over the past century has been manifested by an increase in global mean (near) surface (air) temperature (GMST) of about 0.8 K, and by changes in the altitudinal and geographical distributions of air temperature, widespread melting of snow and of land and sea ice, and rising sea level (Solomon et al. 2007). This warming has been ascribed by the Intergovernmental Panel on Climate Change (IPCC) as "very likely" due mainly to increased concentrations of long-lived greenhouse gases (GHGs), carbon dioxide, methane, chlorofluorocarbons, and nitrous oxide. Confidence in the quantitative attribution of these changes to the products of human activity is essential to planning and decision making regarding global energy resources. However, the observed increase of GMST over the industrial period is less than 40% of what would be expected from present best estimates of the earth's climate sensitivity and the forcing (imposed change in energy balance,

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W m$^{-2}$) by the observed increases in GHGs. Here we examine the discrepancy between the observed increase in GMST over the industrial period and that expected from the increased GHG concentrations and four major factors that might contribute to this discrepancy: (i) natural variation in global temperature over the industrial period, (ii) lack of attainment of equilibrium of the climate system to applied forcings over the industrial period, (iii) current estimates of climate sensitivity being too high, and (iv) countervailing forcings over the industrial period offsetting the warming forcings by incremental greenhouse gases. We show that relatively little of this warming discrepancy can be attributed to a countervailing natural cooling over this time period or to thermal lag of the climate system response to forcing. We argue that this discrepancy is therefore due mainly to offsetting forcing by increased concentrations of atmospheric aerosols and/or to climate sensitivity being lower than current estimates; the discrepancy cannot be apportioned between these two causes primarily because of present uncertainty in aerosol forcing.

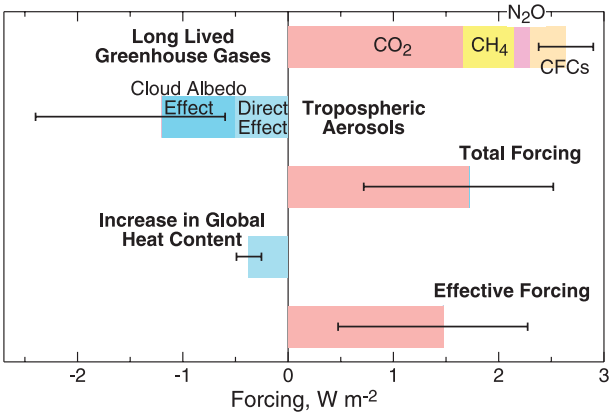


FIG. 1. Global average radiative forcing estimates and associated uncertainty ranges (5%–95% confidence intervals, roughly 1.6 standard deviation) in 2005, relative to the preindustrial climate, for anthropogenic aerosols and long-lived greenhouse gases as given by the IPCC Assessment Report (Solomon et al. 2007). Total forcing includes other anthropogenic and natural (solar) forcings not shown, of which forcing by tropospheric ozone (0.35 W m$^{-2}$) is the greatest. Effective forcing is equal to total forcing minus average increase in global heat content, which is expressed as a flux and shown as a negative quantity.

## 2. Discrepancy between observed and expected temperature increase

According to the forcing-response paradigm that underlies virtually all interpretation of climate change over the industrial period, imposition of a sustained positive (warming) perturbation (forcing, $F$) to the radiation budget of a system initially in radiative equilibrium results in heating of the system (rate of increase in global heat content $H$, $N \equiv dH/dt$) and an increase in GMST, $\Delta T$. The increase in temperature results in an increase in the outgoing thermal infrared radiation, which to first order is proportional to the increase in temperature, yielding for the heating rate:

$$N = F - S^{-1}\Delta T, \tag{1}$$

where $S^{-1}$ is a proportionality constant. When a new radiative equilibrium is reached, $N = 0$ and the equilibrium increase in GMST is

$$\Delta T_{\text{eq}} = SF, \tag{2}$$

from which $S$ is seen to be the equilibrium climate sensitivity in systematic units [i.e., K (W m$^{-2}$)$^{-1}$]. The more commonly used measure of climate sensitivity is the so-called $CO_2$ doubling temperature $\Delta T_{2\times}$, the equilibrium temperature increase that would result from a sustained doubling of atmospheric $CO_2$. This quantity is related to $S$ as $\Delta T_{2\times} = F_{2\times}S$, where $F_{2\times}$, the forcing by doubled $CO_2$, is approximately 3.7 W m$^{-2}$. Forcing by incremental concentrations of long-lived GHGs over the industrial period (to 2005) is about 2.6 W m$^{-2}$ (Fig. 1), which is roughly 70% of $F_{2\times}$. Such a forcing, together with the IPCC best estimate of $\Delta T_{2\times}$ (i.e., 3 K), would thus suggest that the increase in GMST should have been about 2.1 K, well in excess of the observed increase (Solomon et al. 2007) of about 0.8 K (Fig. 2). Forcing by incremental tropospheric ozone, estimated as 0.35 W m$^{-2}$ would contribute an additional 0.3 K to the expected increase in GMST, raising this to 2.4 K. Possible reasons for the warming discrepancy are examined here.

### *a. Natural variation in global temperature*

Some or all of the warming expected over the industrial period might have been offset by cooling over this time period due to natural variability of the climate system. We use variation in preindustrial global temperature as inferred from proxy records, mainly tree rings, ice cores, corals, and varved sediments to estimate the likely magnitude of any natural cooling over the 150-yr interval of the instrumental record. The standard deviation of the difference in temperature over 150-yr intervals for the period (1000–1850) based on the synthesis reconstruction of Juckes et al. (2007) yields 0.2 K, which is 25% of the observed increase in GMST (Fig. 2). Somewhat smaller changes in GMST were found in simulations of the twentieth century with coupled ocean–atmosphere global climate models using estimated natural forcings only (as reported by Solomon et al. 2007, see their Fig. 9.5), which for 19 runs with 5 models yielded

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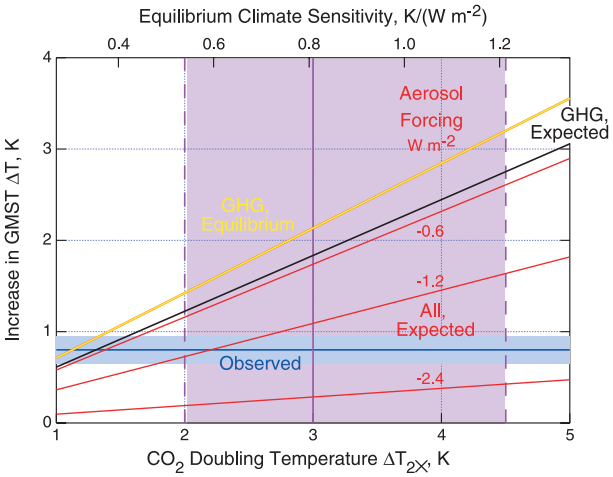


FIG. 2. Equilibrium increase (yellow) and expected present increase (black) in GMST above preindustrial temperature from the forcing by present (2005) incremental concentrations of long-lived greenhouse gases as a function of $CO_2$ doubling temperature $\Delta T_{2\times}$, (bottom axis) or equilibrium climate sensitivity $S$ (top axis); expected increase accounts for global heating rate. Vertical purple solid and dashed lines denote the IPCC best estimate and central 66% uncertainty range for equilibrium climate sensitivity. Red lines show expected increase in GMST, accounting for all forcings over the industrial period (Solomon et al. 2007, their Fig. SPM-2), for indicated values of forcings by anthropogenic aerosols. The blue line denotes observed increase in GMST of present climate relative to preindustrial (0.8 K); the blue band denotes estimated portion of increase in GMST that can be attributed to natural variability.

a temperature increase of 0.09 K (standard deviation is 0.19 K, maximum is 0.49 K). An offset of expected warming by a natural cooling of 0.2 K would account only for about 15% of the discrepancy between observed increase in GMST and that expected for the IPCC's best-estimate sensitivity, $\Delta T_{2\times} = 3$ K (8%, 32%, for $\Delta T_{2\times} =$ 4.5 K, 2 K, respectively). If, on the other hand, natural variability has contributed to increase GMST over the industrial period, the discrepancy would be even greater, by similar amounts.

### b. *Lack of attainment of equilibrium*

The effect of the increase in heat content of the climate system that would be due to disequilibrium, that is, the lack of attainment of steady state subsequent to imposition of a forcing, can be examined empirically. From Eq. (1)

$$S = \frac{\Delta T}{F_{\text{eff}}}, \tag{3}$$

where

$$F_{\text{eff}} \equiv F_{\text{tot}} - N \tag{4}$$

is an effective forcing that is less than the total imposed forcing $F_{\text{tot}}$ by the heat flux into the planet (Cubasch et al. 2001; Gregory et al. 2002). This heat flux is manifested mainly in heating of the World Ocean. Here we estimate $N$, expressed as watts per square meter of the earth's surface, from the rate of increase in ocean heat content from the surface to 700 m (Levitus et al. 2009) using the mean for the time periods 1955–2008 and 1969–2008, with the uncertainty taken to encompass those two rates and their uncertainties. This heating rate is increased by a factor of 1.44 (Levitus et al. 2005) to account for the heating below this depth (to 3000 m), yielding $0.31 \pm 0.10$ W m$^{-2}$, somewhat greater than the value $0.21 \pm 0.04$ given by Solomon et al. (2007). This heat flux is multiplied by another factor of 1.19 (Levitus et al. 2005) to account for additional, minor heat sinks, mainly heating of the atmosphere and solid earth, and melting of land and sea ice (cf. also Huang 2006), to give $N = 0.37 \pm 0.12$ W m$^{-2}$.

Values given by other investigators (Gouretski and Koltermann 2007; Domingues et al. 2008; Wijffels et al. 2008; Ishii and Kimoto 2009) of the planetary heating rate determined from heat flux into the ocean are for the most part comparable to or less than the value presented here. From the short-term time derivative of the ocean heat content given by Domingues et al., Douglass and Knox (2009) infer a time-dependent planetary heating rate that ranges from $-0.15$ W m$^{-2}$ (i.e., cooling) to $+0.15$ W m$^{-2}$; by a similar analysis Murphy et al. (2009) infer a time-dependent planetary heating rate that varies from $-0.6$ W m$^{-2}$ to $+0.8$ W m$^{-2}$ (1954–2001 average, $0.24 \pm 0.32$ W m$^{-2}$, $1\sigma$). A considerably greater planetary heating rate was given by Willis et al. (2004), based on an indicated heating rate of $0.86 \pm 0.12$ W m$^{-2}$ of the ocean from 1993 to 2003 for the upper 750 m of the water column (0.61 W m$^{-2}$ globally). This heat flux had served as the basis for the conclusion by Hansen et al. (2005) of a substantial planetary radiative imbalance, which in turn served as the basis for the imbalance given in the review of the global energy budget by Trenberth et al. (2009). However, much of the apparent increase in heat content reported by Willis et al. is now attributed to problems in the measurements (Gouretski and Koltermann 2007; Wijffels et al. 2008), so such a large heating rate and planetary radiative imbalance seems unlikely; for these reasons we retain the value of $0.37 \pm 0.12$ W m$^{-2}$.

Because a planetary heating rate of $0.37 \pm 0.12$ W m$^{-2}$ is only a small fraction, $14\% \pm 5\%$, of the forcing by long-lived GHGs (Fig. 1), the heat flux into the planet can account for only a modest reduction in the expected increase in GMST due to forcing by increases in long-lived GHGs, relative to the equilibrium increase in

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GMST (Fig. 2). For sensitivity $\Delta T_{2\times} = 3$ K, the corresponding fraction of the warming discrepancy attributable to thermal disequilibrium is 22% ± 7% (19% ± 6% and 32% ± 10% for $\Delta T_{2\times} = 4.5$ K and 2 K, respectively). Thus, the thermal lag of the climate system accounts for a small, albeit nonnegligible, fraction of the warming discrepancy. A similar result, 10% ± 7%, was obtained by Murphy et al. (2009) by comparing the increase in ocean heat content between 1950 and the present with the integral of the greenhouse gas plus solar forcing over that period.

### c. Overestimate of climate sensitivity

The sensitivity of the earth's climate system is thought to substantially exceed that of a blackbody radiator at the global mean surface temperature 288 K, $\Delta T_{2\times} = 1.1$ K, because of feedbacks in the climate system. These feedbacks are not well understood, with the consequence that the climate sensitivity is quite uncertain. The IPCC Assessment Report (Solomon et al. 2007) specified the ''likely'' uncertainty range associated with its estimate of climate sensitivity by the central 66% of the probability distribution function characterizing this quantity (2.0–4.5 K; Fig. 2). The IPCC report further stated that it was ''very unlikely'' (less than 5% probability) that the climate sensitivity is less than 1.5 K, but was unable to recommend a corresponding very unlikely upper bound to the estimate, stating rather that on the basis of present understanding values greater than 4.5 K could not be excluded. The expected increase in GMST that would result from forcing by only the long-lived GHGs is well above the observed increase (Fig. 2) for the entire likely range in climate sensitivity given by the IPCC. Consequently the actual sensitivity must be even lower than the IPCC very unlikely limit if other causes are negligible; alternatively, the low observed increase in GMST must be attributed at least in part to other causes.

### d. Countervailing forcings over the industrial period

Tropospheric aerosols from human activity scatter light (direct effect) and increase cloud reflectivity (the cloud albedo effect) and possibly also cloud persistence; all these effects are thought to exert a negative forcing (cooling influence) on climate. Some aerosols absorb light and can exert a positive forcing, and can also decrease the persistence of clouds by enhancing evaporation of droplets and suppressing convection (Ramanathan and Feng 2008). The combined positive and negative aerosol forcings yield the IPCC best estimate of $-1.2$ W m$^{-2}$ (5%–95% range $-0.6$ to $-2.4$ W m$^{-2}$) that is substantial compared to the forcing by long-lived GHGs but quite uncertain (Fig. 1). Under the assumption of additivity of forcings (a premise of the linear forcing-response paradigm), the best estimate of the total aerosol forcing, in global and annual average, would be offsetting 45% of the forcing by the long-lived GHGs. An aerosol forcing of $-1.2$ W m$^{-2}$ would reduce the effective forcing over the industrial period to 1.5 W m$^{-2}$. This would result in a much lower expected increase in GMST than would forcing by GHGs alone (Fig. 2) and would actually be compatible with the lower end of the IPCC likely range of climate sensitivity. However, the large uncertainty associated with present estimates of aerosol forcing and the resultant uncertainty in the total effective forcing over the industrial period (Fig. 1) imply a range of expected increase in GMST that is compatible with, and extends well beyond, the entire range of the 2007 IPCC estimated climate sensitivity. Thus, countervailing aerosol forcing could account for much or all of the discrepancy between the expected and observed increase in GMST over the industrial period.

## 3. Determining climate sensitivity

In view of the central role of climate sensitivity in understanding and projecting the response of the earth's climate to future changes in atmospheric composition, much effort has been directed toward determining this quantity, or at least bounding it. Fundamentally there are only two generally accepted approaches to determining climate sensitivity: *empirical* and *modeling*.

### a. Empirical approaches

Empirical approaches consist of attributing and quantitatively relating an observed change in GMST over some time period to a known or estimated forcing. The underlying assumptions are that the magnitude of the observed temperature change is greater than unforced variability, that there is a cause and effect relationship between the net assigned forcing and the observed temperature change, and that the observed change reflects the full impact of the forcing or that disequilibrium can be accounted for. Knowledge of both the responsible forcing and the resulting temperature change is required. Prehistoric periods of change in GMST to which this approach has been applied include the warming between the last glacial maximum (LGM) and the present temperate era and cooling between the Cretaceous period and the present; changes in global temperature are inferred from changes in the distribution of stable isotopes in ice cores and in sediments, and changes in atmospheric composition are inferred from ice cores or from mass balance considerations. Although this paleological approach is thought by many to give reliable estimates of the earth's climate sensitivity, the uncertainties

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are substantial. The IPCC report (Solomon et al. 2007, chapter 6) gives the total forcing between the present and the LGM as "approximately $-8$ W m$^{-2}$," with no uncertainty range specified but with the level of scientific understanding ascribed to the main non-GHG forcings, which arise from changes in continental ice and sea level, mineral dust, and vegetation ($-3.2$, $-1.4$, and $-1.2$ W m$^{-2}$, respectively), indicated as low, very low, and very low, respectively. The very likely range of change in GMST is given as $-4$ to $-7$ K. The range of climate sensitivity ($\Delta T_{2\times}$) due to the uncertainty in temperature change alone is 1.9–3.2 K, but this uncertainty range is surely an underestimate because of uncertainty in the forcing. Based on estimated uncertainties in the magnitude of temperature change in the Antarctic between the present and the LGM ($-7$ to $-11$ K), the ratio of the Antarctic temperature change to global mean change (1–2), and the forcing between the two climate states ($-6$ to $-10$ W m$^{-2}$); Ganopolski and Schneider von Deimling (2008) conclude that the range of $\Delta T_{2\times}$ that can be inferred from this transition cannot be narrowed below 1.3–6.8 K. A further concern would be the applicability of the climate sensitivity inferred from such a large forcing and temperature change to the smaller anthropogenic perturbations associated with response to forcing by incremental GHGs.

Another situation of known forcing and temperature response that might in principle be used to infer equilibrium climate sensitivity is cooling following an explosive volcanic eruption. However, because of the short duration of the forcing, this sensitivity cannot be inferred from the temperature record unless the rate of change of global heat content is known explicitly (Boer et al. 2007).

A potentially very valuable empirical approach is to determine climate sensitivity from the known forcing and the increase in GMST over the industrial period, accounting for disequilibrium by the global heating rate (Gregory et al. 2002). For the increase in GMST taken as 0.8 K and the effective forcing evaluated by Eq. (4) with the IPCC Fourth Assessment Report (AR4) best estimate of forcing over the industrial period, $F_{\rm tot} = 1.72$ W m$^{-2}$, and planetary heating rate $N = 0.37$ W m$^{-2}$, the resulting sensitivity, evaluated as $S = \Delta T/F_{\rm eff}$ is 0.59 K (W m$^{-2}$)$^{-1}$, corresponding to a $CO_2$ doubling temperature $\Delta T_{2\times} = 2.2$ K, near the low end of the likely range of sensitivity given by the IPCC Assessment Report (Solomon et al. 2007). This calculation is illustrated in Fig. 3, a graph of effective forcing versus the inverse of climate sensitivity, for which the known slope (temperature increase over the industrial period) permits sensitivity to be determined for a specified effective forcing. However, as shown in the figure, the present uncertainty in effective forcing, arising mainly from the uncertainty in aerosol forcing, introduces a very large uncertainty into the calculated sensitivity and $CO_2$ doubling temperature, with bounds that exceed the Solomon et al. (2007) likely range at both ends. Further uncertainties are introduced by uncertainty in the forced temperature change over the industrial period, shown by blue shading in the figure, and by uncertainty in forcings other than by aerosols and in the heating rate, estimated to be about $\pm 0.4$ W m$^{-2}$ in the aggregate. However, the dominant source of uncertainty is in the aerosol forcing, and hence any empirical determination of the earth's climate sensitivity from temperature change over the instrumental record is forestalled by the present uncertainty in aerosol forcing.

A variant of this approach is to determine the global heating rate $N$ not from the rate of increase in ocean heat content but from the energy imbalance at the top of the atmosphere (TOA), measured from a satellite, as

$$N = J_S/4 - J^{\uparrow}_{\rm sw} - J^{\uparrow}_{\rm lw}. \tag{5}$$

Here $J_S$ is the solar "constant," the solar irradiance normal to the sun–earth vector at the TOA, a function of earth–sun distance and any variation in solar output, and $J^{\uparrow}_{\rm sw}$ and $J^{\uparrow}_{\rm lw}$ denote upwelling global mean short- and longwave irradiance, respectively; all fluxes are taken as positive. Introducing this heating rate into Eq. (1) yields

$$F - (J_S/4 - J^{\uparrow}_{\rm sw} - J^{\uparrow}_{\rm lw}) = S^{-1}\Delta T, \tag{6}$$

where $F$ is restricted to a forcing below the top of the atmosphere, any forcing due to change in solar irradiance being accounted for by $\Delta J_S/4$. From Eq. (6) the inverse sensitivity $S^{-1}$ is obtained as the slope of a graph of $F - (J_S/4 - J^{\uparrow}_{\rm sw} - J^{\uparrow}_{\rm lw})$ versus $\Delta T$. As with the other empirical approaches, determination of the sensitivity requires knowledge of the forcing.

Such an approach has been applied by Forster and Gregory (2006) and by Murphy et al. (2009) using data from the Earth Radiation Budget Experiment (ERBE) satellite and/or the Clouds and the Earth's Radiant Energy System (CERES) satellites, who obtained much narrower uncertainty bounds than those obtained elsewhere. [Murphy et al. (2009) distinguish the quantity they obtain from the equilibrium climate response.] However, several concerns might be noted. Both studies assumed anthropogenic aerosol forcing to be unchanged during the relatively short durations of the satellite measurements examined [ERBE data 1985–97, Forster and Gregory (2006); ERBE data 1985–99 and CERES data 2000–05, Murphy et al. (2009)], without examining the consequences of this assumption. A second concern is that the changes in annual mean temperature and radiative

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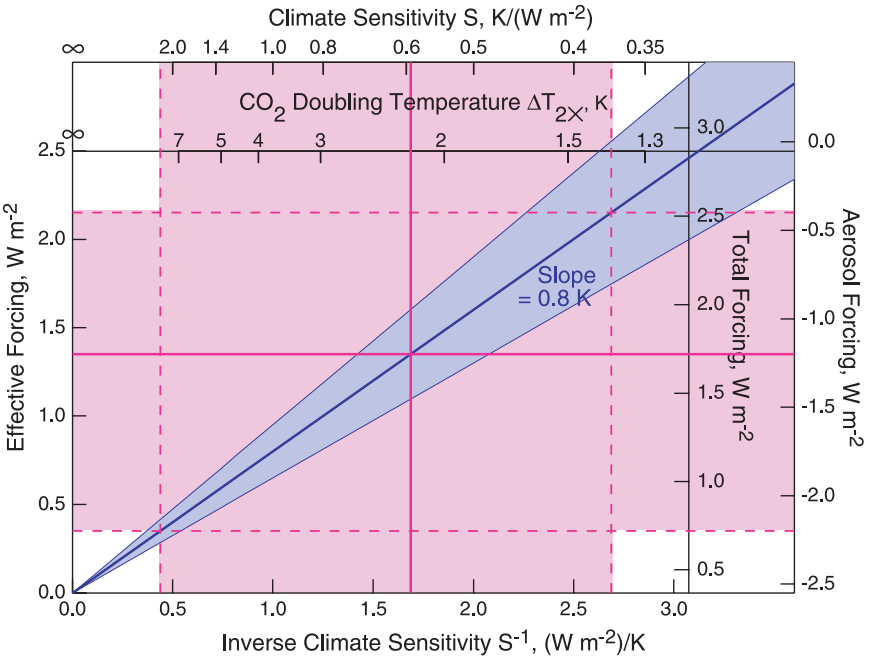


FIG. 3. Empirical determination of the earth's climate sensitivity. The blue line denotes the relation between climate sensitivity $S$ top axis or $CO_2$ doubling temperature $\Delta T_{2\times}$, auxiliary top axis, both on inverse scale, and effective forcing over the industrial period, left axis; total forcing, right auxiliary axis; or aerosol forcing, right axis, for increase in GMST over industrial period equal to 0.8 K; the blue band notes associated uncertainty due to natural variability. The solid horizontal magenta line denotes the IPCC best estimate (Solomon et al. 2007) of aerosol forcing and total forcing over the industrial period $F_{tot}$, and effective forcing in Fig. 1. Vertical solid line denotes inferred climate sensitivity corresponding to this forcing. Dashed lines and fill denote "likely" uncertainty range in forcing (Fig. 1) and resultant uncertainty range in inferred sensitivity.

fluxes over these rather short periods are so small that any unforced variation in temperature or irradiance or unrecognized forcing would exert a large influence on the inferred climate sensitivity; to overcome this limitation Murphy et al. (2009) determined sensitivity using monthly-mean as well as annual-mean temperatures and fluxes. However, as temperature change over the annual cycle does not account for seasonal variations in large-scale circulations, use of monthly mean data cannot yield an accurate climate sensitivity (e.g., Lindzen et al. 1995). Also, the approach using monthly data would require confident estimate of the seasonal variation of all forcings, including aerosol forcing. The best estimate for $S^{-1}$ and the rather narrow associated $1\sigma$ uncertainty range presented by Murphy et al. (2009) from analysis of the seasonal data $(1.25 \pm 0.5)$ W m$^{-2}$ K$^{-1}$ differ markedly from the values they obtained for the three annual datasets examined $(0.04 \pm 1.03, 1.03 \pm 0.58, 0.69 \pm 0.78)$ W m$^{-2}$ K$^{-1}$ and from the value obtained by a similar analysis of a subset of the same data by Forster and Gregory, $2.3 \pm 0.7$ W m$^{-2}$ K$^{-1}$ $(1\sigma)$. A further concern would be the suitability of the satellite data to obtaining the necessary global average fluxes: the ERBE satellite does not sample poleward of 60°; the CERES satellite, although capturing the full planet from polar orbit, does not sample over the diurnal cycle. All of these considerations would seem to limit the confidence that can be placed in determination of climate sensitivity by this approach, at least at present.

### b. *Climate modeling approach*

Global climate models, numerical representations of the processes that comprise the earth's climate system, permit a highly differentiated examination both of prior climate change and of future climate change that would result from prospective changes in atmospheric composition. As such, they are the most powerful tools available for examining the consequences of perturbations of the climate system. Determining climate sensitivity by this approach requires development and evaluation of models that represent the key processes of the climate system with sufficient accuracy that the consequences of the rather small GHG perturbations can be determined accurately and with confidence. Current climate models

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exhibit considerable diversity in their representations of key atmospheric and hydrologic processes, mainly because of insufficient understanding and computational limitations; changing representations of individual physical processes, even within a given climate model, can result in large changes in the model's climate sensitivity (Sanderson et al. 2008), mainly because of the amplifying effects of feedbacks (e.g., Hansen et al. 1984; Schlesinger 1988; Roe and Baker 2007). For reasons such as these, current climate models exhibit a range of sensitivity of almost a factor of 2, roughly equal to the IPCC likely uncertainty range for this quantity. As evaluation of climate models requires comparison with observations, accurate knowledge of all forcings, including aerosol forcings, is required.

The limitations on determination of climate sensitivity imposed by the present uncertainty in forcing by anthropogenic aerosols are increasingly becoming recognized for both the empirical and modeling approaches. Hansen (2008) has argued that ''estimates of climate sensitivity based on the last 100 years of climate change are practically worthless, because we do not know the net climate forcing.'' Uncertainty in aerosol forcing allows the observed change in GMST over the twentieth century, to be reproduced with climate models exhibiting a wide range of sensitivities (Schwartz et al. 2007); this has been ascribed in large part to an inverse correlation of the forcings chosen and employed by the several modeling groups with the sensitivities of their models (Kiehl 2007). This situation led Knutti (2008) to conclude recently that ''constraining the aerosol effects from data, models and from the observed warming trends is therefore a critical step in order to decide whether our understanding of human influence on climate and our climate models are consistent with observed trends.'' Clearly, progress in reducing uncertainty in climate sensitivity by either the empirical or modeling approach requires reducing the uncertainty in aerosol forcing. Even if the modeling approach converged upon a narrow range for climate sensitivity, it would still be essential to reduce the uncertainty in forcing to test the models' ability to reproduce the twentieth-century temperature record, in order to place confidence in the models. For these reasons, reducing uncertainty in aerosol forcing is of singular importance for reducing uncertainty in climate sensitivity.

## 4. Strategies for reducing uncertainty in aerosol forcing

Over the three plus decades since the initial attempts to identify and quantify the direct and cloud albedo climate forcings by anthropogenic aerosols (Junge 1975), major advances have been made in characterizing aerosol properties and distributions and in understanding the controlling processes (Climate Change Science Program 2009). Recognition of the need for a global system for observing anthropogenic aerosols has led to the development and deployment of substantial capabilities for remote sensing of aerosols from satellites and the surface (Kahn et al. 2004; Yu et al. 2006). Key current satellite capabilities from instruments orbiting in formation in low earth orbit yield global near-simultaneous observations of radiance from aerosols at multiple wavelengths and scattering angles and of vertical profiles by lidar (Anderson et al. 2005). Advanced satellite instruments promising additional, urgently needed capabilities are being readied for launch in the near future (e.g., the Glory Mission; Mishchenko et al. 2007), or, more importantly, could be readied for early deployment as in the Deep Space Climate Observatory (DSCOVR), which would be stationed at the Lagrange point between the sun and earth to provide continuous calibrated measurements of clouds and aerosols (Yeager 2008). Surface-based networks yield multiwavelength aerosol optical depth and angular scattering distribution by sun and sky photometry and vertical distributions by lidar at multiple aerosol-impacted locations around the globe. Near-surface in situ observations can provide very detailed long-term but spatially limited coverage of aerosol optical, microphysical, and chemical properties (Delene and Ogren 2002; Malm et al. 2004; Zhang et al. 2007). Together these measurements provide essential information about the abundance and properties of natural and anthropogenic aerosols as well as ground truth information needed to evaluate the satellite aerosol retrievals. A dozen or so major short-term field studies in a variety of locations globally have provided detailed coincident measurements of multiple environmental attributes related to key aerosol processes (e.g., new particle formation and gas-to-particle conversion rates) and properties (e.g., size distributed chemical composition for identifying the anthropogenic fraction; Bates et al. 2006). Substantial advances are being made as well in modeling the global distribution of aerosol amounts and properties, the geographical distribution of aerosol influences, and the attribution to sources and in intercomparing the model results (Kinne et al. 2006). Such modeling, which requires input and constraints from process studies, in situ measurements, and satellite observations, is receiving increased attention by the research community and is making much progress (Climate Change Science Program 2009).

Despite these many advances, the ability to model the amount, properties, and geographical distribution of anthropogenic aerosols with the accuracy required to confidently calculate their radiative forcing either in

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stand-alone mode or in climate models has thus far been elusive (Ghan and Schwartz 2007). Also, and importantly, essential observational data are not available to directly yield forcing or to adequately constrain global aerosol models. Satellite measurements alone are unable to reduce the uncertainty in aerosol forcing because of sampling limitations and biases (Kahn et al. 2009) and currently cannot unambiguously attribute satellite-derived quantities to natural versus anthropogenic aerosols (Anderson et al. 2005); also, satellite measurements are limited in their ability to detect the smallest-sized aerosol particles that influence clouds (Heintzenberg and Charlson 2009). These considerations suggest that a combination of approaches is required to make progress in quantifying aerosol forcing. This quantification and its relationship to anthropogenic emissions would be greatly advanced by the ability to integrate global measurements of aerosol and cloud properties into radiative transfer models (e.g., in a reanalysis mode). Such a linking of satellite remote sensing, in situ observations, and models, which has been called for repeatedly (Charlson et al. 1992; National Research Council 1996, 2005; Diner et al. 2004; Climate Change Science Program 2009) could be expected to make rapid progress. Although the key elements of the needed research program are present, progress has been slow because of the lack of a coherent international or even national program to bring these elements together. A focused, integrated approach could lead to the understanding of aerosol processes needed to represent them in chemical transport models and climate models with the required accuracy in forcing. What seems to be absent is the integration of the necessary observational and modeling capabilities.

## 5. Implications

The preceding analysis establishes that the warming discrepancy is due to some combination of low climate sensitivity and/or offset of the expected increase in GMST due to incremental GHG concentrations by other forcing influences, of which the most likely candidate is anthropogenic aerosols. As shown below, the actual situation and the extent of aerosol offset are of enormous importance in understanding climate change over the industrial period and in developing policy to deal with future climate change. Importantly, the present large uncertainty in climate sensitivity results in an even greater uncertainty in the amount of additional carbon dioxide that can be added to the atmosphere for a given allowable increase in global mean temperature.

The implications of the present uncertainty in climate sensitivity on allowable future $CO_2$ emission can be readily quantified by a simple calculation. For a given climate sensitivity $S$ and target maximum increase in GMST above preindustrial not to be exceeded, $\Delta T_{\max}$, it is possible to calculate the total further amount of equivalent $CO_2$, in addition to the incremental amounts of $CO_2$ and other long-lived GHGs in the present atmosphere above their preindustrial values, that can be introduced into the atmosphere; here the term "equivalent $CO_2$" includes other long-lived GHGs with emissions scaled to that of $CO_2$ by their specific forcings. An example of such a calculation is given in Table 1 for the target maximum increase in GMST $\Delta T_{\max}$ taken as 2 K, a widely adopted value proposed by the European Union as early as 1996 based on consideration of risk (European Union Council 1996). The choice of $\Delta T_{\max}$ is ultimately a societal decision that must take into consideration the consequences of a given increase in GMST, the costs of achieving such a temperature stabilization, and the uncertainties in the calculation; results for other values of $\Delta T_{\max}$ are shown in Fig. 4. The calculation is made here for the best-estimate value for $\Delta T_{2\times}$ as given by the 2007 IPCC Assessment Report (3 K) and for the values corresponding to the limits of the very likely and likely uncertainty range–cumulative values of the probability distribution function for this quantity of 5% (1.5 K, roughly $1.6\sigma$ below the best estimate), 17% (2 K, $1\sigma$ below), and 83% (4.5 K, $1\sigma$ above); as noted above the IPCC report did not provide an estimate of $\Delta T_{2\times}$ for 95% cumulative probability. In this calculation it is necessary to account for the forcing by incremental amounts of $CO_2$ and other long-lived GHGs already in the atmosphere. The calculation does not take into account any disequilibrium contribution, which would result only in a time lag in the temperature increase. The calculation excludes the cooling influence of anthropogenic aerosols because of their short atmospheric residence times, about a week. Similarly, the warming forcing due to increases of tropospheric ozone is excluded as this short-lived GHG would likely decrease rapidly as emissions of precursors, mainly nitrogen oxides associated with fossil fuel combustion, were decreased in response to future pollution controls or as future $CO_2$ emissions were decreased. Additionally, the table presents the time in years allowed for total $CO_2$ emission to continue at the present rate of $CO_2$ emission from fossil fuel combustion and cement manufacture that would correspond to the allowable cumulative emission.

This simple calculation (for more detailed calculations see Caldeira et al. 2003; Edmonds and Smith 2006; Allen et al. 2009) illustrates the dramatic differences in future fuel use scenarios between low and high climate sensitivity. For climate sensitivity at the low end of the very likely range of the 2007 IPCC estimate

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TABLE 1. Allowable future equivalent $CO_2$ emission* for increase in GMST above its preindustrial value not to exceed $\Delta T_{max} = 2$ K.

| Quantity | Symbol | Unit | Value | | | |
|---|---|---|---|---|---|---|
| $CO_2$ doubling temperature at equilibrium | $\Delta T_{2\times}$ | K | 1.5 | 2 | 3 | 4.5 |
| Equilibrium climate sensitivity | $S$ | K (W m$^{-2}$)$^{-1}$ | 0.40 | 0.54 | 0.81 | 1.21 |
| Cumulative probability that actual doubling temperature $> \Delta T_{2\times}$ | $P(\Delta T_{2\times})$ | % | 5 | 17 | ~50 | 83 |
| Expected current equilibrium increase in GMST for indicated doubling temperature | $\Delta T_c$ | K | 1.1 | 1.4 | 2.1 | 3.2 |
| Allowable future increase in GMST | $\Delta T_a$ | K | 0.9 | 0.6 | −0.1 | −1.2 |
| Allowable future increase in $CO_2$ mixing ratio | $\Delta m_{CO_2}$ | ppm | 164 | 76 | −12 | −70 |
| Target $CO_2$ mixing ratio | $m_{CO_2}$ | ppm | 544 | 456 | 368 | 310 |
| Allowable cumulative future $CO_2$ emission | $E_{CO_2}$ | Pg C | 697 | 323 | −50 | −299 |
| Time at present $CO_2$ emission rate to reach $\Delta m_{CO_2}$ | $t_{CO_2}$ | yr | 77 | 36 | −6 | −33 |

* The allowable incremental mixing ratio of equivalent atmospheric $CO_2$ above present that is compatible with a target maximum temperature increase above preindustrial temperature $\Delta T_{max}$, taken here as 2 K, is evaluated as $\Delta m_{CO_2} = (\Delta T_{max} - \Delta T_c)/Sf = \Delta T_{max}/Sf - F_c/f$ where $\Delta T_c$ is the equilibrium increase in GMST that would be expected from incremental long-lived gases above preindustrial in the current atmosphere, and $F_c = 2.6$ W m$^{-2}$ is the corresponding forcing; $S$, the equilibrium climate sensitivity in units of K/(W m$^{-2}$), is related to $CO_2$ doubling temperature at equilibrium $\Delta T_{2\times}$ as $S = \Delta T_{2\times}/F_{2\times}$ where $F_{2\times} = 3.7$ W m$^{-2}$ is the forcing for doubled $CO_2$; and $f = 0.0141$ W m$^{-2}$ ppm$^{-1}$ is the specific forcing, i.e., the forcing per incremental ppm of $CO_2$, evaluated as $f \equiv F/\Delta m_{CO_2} \approx F_{2\times}/(m_c \ln 2)$ where $m_c$ is the current atmospheric $CO_2$ mixing ratio, 380 ppm. The corresponding allowable cumulative future emissions of long-lived greenhouse gases, expressed as equivalent $CO_2$, is evaluated as $E_{CO_2} = \Delta m_{CO_2}/cr$ where $c$ is a conversion factor between $CO_2$ emission and atmospheric mixing ratio, 0.47 ppm/Pg C, and $r$ is the fraction of emitted $CO_2$ that remains in the atmosphere, taken here as 0.5. The time remaining until the cumulative allowable amount of additional $CO_2$ would be reached at the present rate of emission of $CO_2$ from fossil fuel combustion and cement production $q \approx 9$ Pg C yr$^{-1}$ (Raupach et al. 2007) is evaluated as $t_{CO_2} = E_{CO_2}/q$; a negative value indicates that the present atmospheric $CO_2$ mixing ratio 380 ppm exceeds the allowable value by an amount that corresponds to the indicated number of years at the present emission rate. Calculations are presented for $\Delta T_{2\times} = 3$ K, the best estimate for this quantity given by the Solomon et al. 2007 Assessment Report, and for $\Delta T_{2\times} = 1.5$, 2, and 4.5 K., corresponding to cumulative probability for this quantity given in that report $P(\Delta T_{2\times}) = 5$, 17, and 83%. The calculation neglects cooling due to forcing by aerosols and warming due to forcing by ozone, as discussed in text. 1 Pg = 1 gigatonne = $10^{15}$ g.

($\Delta T_{2\times} = 1.5$ K), the cumulative allowable future emission of equivalent $CO_2$ compatible with a maximum increase in GMST above preindustrial $\Delta T_{max} = 2$ K is about 700 Pg C; at the present rate of emission of $CO_2$ from fossil fuel combustion and cement manufacture ~9 Pg C yr$^{-1}$ (Raupach et al. 2007), this allowable increase in $CO_2$ would be reached in about 70 yr. For a somewhat greater sensitivity $\Delta T_{2\times} = 2$ K, corresponding to the low end of the IPCC likely range, the cumulative future allowable equivalent emission of $CO_2$ is a factor of 2 lower, and the allowable cumulative emission would be reached in about 40 yr. For the IPCC best estimate of climate sensitivity ($\Delta T_{2\times} = 3$ K) the present committed increase in GMST due to incremental GHGs already present in the atmosphere (2.1 K) essentially already equals the target maximum 2-K increase in GMST above preindustrial. On the basis of present understanding of climate sensitivity this possibility cannot be precluded. Thus, if the $CO_2$ doubling temperature of the earth's climate is 3 K, an immediate cessation of emission of $CO_2$ and other GHGs would be required for the equilibrium temperature increase above preindustrial not to exceed 2 K. If climate sensitivity is 4.5 K, at the high end of the IPCC estimated range, the present committed temperature increase above preindustrial is 3.2 K, which substantially exceeds the target maximum temperature increase of 2 K, and attaining the target maximum temperature increase would require substantial reduction in atmospheric GHG concentrations from their present values. The possibility of this situation also cannot be precluded on the basis of present understanding.

In principle, a greater amount of incremental $CO_2$ might be allowed if the cooling influence by aerosols were accounted for in the calculation. However, reliance on continued aerosol offset of warming by GHGs would ultimately fail, as the short residence time of these aerosols would entail a sustained commitment to replenish aerosol concentrations, even as $CO_2$ emissions were reduced. Climate model calculations (e.g., Brasseur and Roeckner 2005; Matthews and Caldeira 2007) find that abruptly decreasing aerosol forcing would result in a rapid jump in temperature as the climate system reacts to the increased net forcing.

The strong dependence of allowable future $CO_2$ emissions on climate sensitivity, which holds for other choices of maximum allowable increase in GMST (Fig. 4) underscores the need to improve knowledge of the earth's climate sensitivity for understanding climate change that has already occurred and for developing strategies to limit future increase in GMST.

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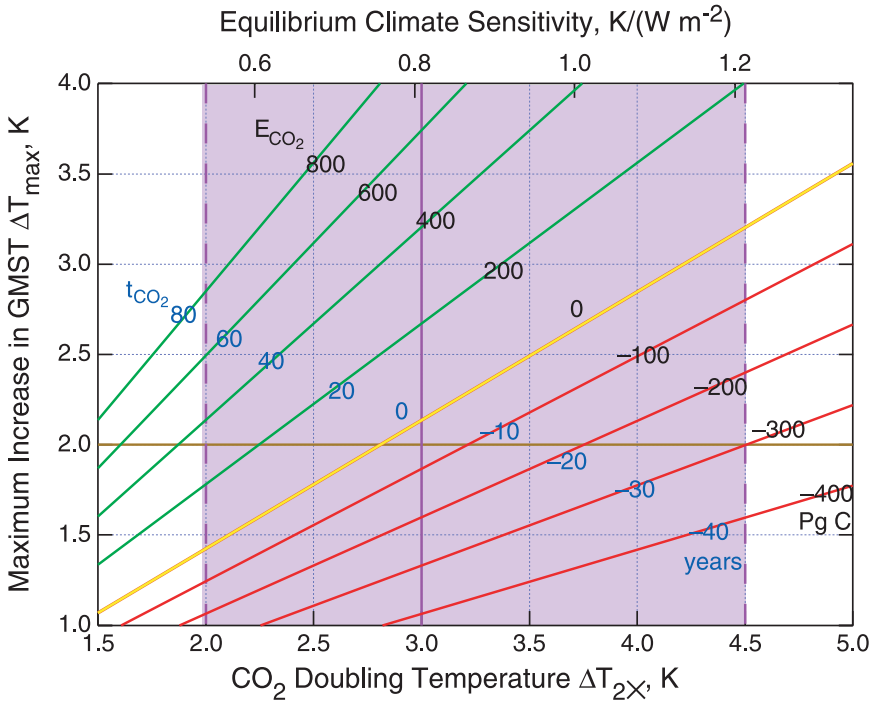


FIG. 4. Allowable cumulative future equivalent $CO_2$ emission $E_{CO_2}$, Pg C (green), compatible with a given maximum acceptable increase in GMST above preindustrial $\Delta T_{max}$ (ordinate) as a function of equilibrium climate sensitivity expressed as $CO_2$ doubling temperature $\Delta T_{2\times}$, (abscissa). Line $E_{CO_2} = 0$ (yellow) denotes expected equilibrium increase in GMST above preindustrial for present incremental greenhouse gas concentrations as a function of $\Delta T_{2\times}$ as in Fig. 2. Negative values of $E_{CO_2}$ (red) indicate amount of equivalent $CO_2$ that would need to be removed from present atmosphere to achieve a given $\Delta T_{max}$ as a function of $\Delta T_{2\times}$. Time (years) corresponding to cumulative emission $E_{CO_2}$ is evaluated as $t_{CO_2} = E_{CO_2}/q$, where the current emission rate $q$ is taken for simplicity as 10 Pg C $yr^{-1}$. Vertical purple line indicates the best estimate and the shaded area denotes range (66% likelihood) for $CO_2$ doubling temperature given by the 2007 IPCC Assessment Report, as in Fig. 2. Horizontal brown line indicates widely accepted value of maximum allowable increase in GMST. Calculations do not account for the cooling influence of enhanced short-lived atmospheric aerosols or the warming influence of enhanced tropospheric ozone.

## 6. Conclusions

The current best estimate and uncertainty range of the earth's climate sensitivity suggest an equilibrium increase in the earth's global mean surface temperature for forcing by anthropogenic long-lived greenhouse gases of 2.1 K (range 1.5–3.2 K, roughly 1 standard deviation), well in excess of the observed increase relative to preindustrial times, about 0.8 K. The discrepancy is attributed mainly to uncertainty in climate sensitivity and/or cooling forcing by anthropogenic aerosols, also highly uncertain; countervailing natural cooling and thermal lag in climate response seem to be relatively small. Because of the great difference in atmospheric residence times of greenhouse gases and aerosols, the effect of the greenhouse gases will dominate long-term forcing and climate response. Even if the earth's climate sensitivity is at the low end of the IPCC estimated "likely" range, continued emission of $CO_2$ at the present rate would exhaust in just a few decades the shared global resource of the incremental amount of $CO_2$ that can be added to the atmosphere without exceeding proposed maximum increases in GMST. If the sensitivity is greater, the allowable incremental emission decreases sharply, essentially to zero at the present best estimate of climate sensitivity, and is actually negative for greater values of this sensitivity. As has been widely discussed elsewhere, redirecting the world's energy economy from its present reliance on fossil fuels or developing effective means of sequestering $CO_2$ emissions would require immense and rapid changes in how the world meets its energy needs. Advance knowledge of the earth's climate sensitivity would be of enormous monetary value, estimated in the tens of trillions of dollars (Edmonds and

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Smith 2006) in terms of efficient planning and averting costs associated with abandoning fossil fuel plants or retrofitting $CO_2$ sequestration systems to existing facilities. Consequently much improved knowledge of the earth's climate sensitivity is urgently and rapidly required for determining the extent and timing of reductions in $CO_2$ emissions needed to limit the increase in GMST to a given value. The principal limitation to empirical determination of climate sensitivity or to the evaluation of the performance of climate models over the period of instrumental measurements is the present uncertainty in forcing by anthropogenic aerosols. This situation calls for greatly enhanced efforts to reduce this uncertainty.

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