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Uncertainty Requirements in Radiative Forcing of Climate Change

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ABSTRACT

The continuing increase in atmospheric carbon dioxide (CO_2) makes it essential that climate sensitivity, the equilibrium change in global mean surface temperature that would result from a given radiative forcing, be quantified with known uncertainty. Present estimates are quite uncertain, 3 ± 1.5 K for doubling of CO_2. Model studies examining climate response to forcing by greenhouse gases and aerosols exhibit large differences in sensitivities and imposed aerosol forcings that raise questions regarding claims of their having reproduced observed large-scale changes in surface temperature over the 20th century. Present uncertainty in forcing, caused largely by uncertainty in forcing by aerosols, precludes meaningful model evaluation by comparison with observed global temperature change or empirical determination of climate sensitivity. Uncertainty in aerosol forcing must be reduced at least three-fold for uncertainty in climate sensitivity to be meaningfully reduced and bounded.

INTRODUCTION

The sensitivity of global mean temperature change to an increase in atmospheric carbon dioxide (CO_2) is not well established. The complexity of the climate system precludes calculation of the response of Earth's climate to a change in a radiative flux component (forcing[1]) from well-established physical laws. Consequently, determination of global climate sensitivity is a subject of intense research. This work is reviewed from time to time by pertinent national and international bodies. One such landmark review was that of a 1979 National Research Council panel chaired by Charney,[2] which concluded: "We estimate the most probable global warming for a doubling of CO_2 to be near 3 °C, with a probable error of ±1.5°." More recently, the Intergovernmental Panel on Climate Change (IPCC)[3] concluded that "Climate sensitivity [to CO_2 doubling] is likely to be in the range 1.5–4.5 °C." These estimates must be considered somewhat subjective. They are based mainly on calculations with climate models constrained, especially for the IPCC estimate, by observation of the extent of warming over the industrial period and concurrence of modeled and observed warming. Neither the Charney panel nor the IPCC quantitatively specified the meaning of their uncertainty bounds (e.g., 2σ), but in the case of the Charney estimate, a National Research Council panel[4] three years later expressed its understanding that "the Charney group meant to imply a 50% probability that the true value would lie within the stated range." Remarkably, despite some two decades of intervening work, neither the central value nor the uncertainty range has changed.

The large uncertainty range, a factor of 3, in present estimates of climate sensitivity renders such estimates not particularly useful from the perspective of developing policy regarding either reduction of greenhouse gas (GHG) emissions or adaptation to a new, increasingly warm climate. While some comfort might be taken in a sensitivity at the low end of the range $\Delta T_{2\times}$ = 1.5 K (where $\Delta T_{2\times}$ is the equilibrium increase of mean surface temperature that would result from a doubling of atmospheric CO_2; see Appendix), a sensitivity at the high end of the range would, for a doubled-CO_2 atmosphere, which is expected to occur well within this century, result in an overwhelming change in Earth's climate. In this context, it might be noted that the increase in global mean temperature from the middle of the last ice age to the present interglacial age is estimated as 5 or 6 K.[5]

These considerations speak to the need to decrease the uncertainty associated with the climate sensitivity. There are two independent approaches to doing this, by using climate models and by empirical inference from the change in

IMPLICATIONS

The present estimate of the global average climate sensitivity, 3 ± 1.5 K for doubling of CO_2, is too uncertain to usefully inform policy-making regarding mitigation of greenhouse warming or adaptation to a warmer future climate. Approaches to determining this sensitivity, either through use of climate models or by empirical inference, are limited at present by uncertainty in radiative forcing of climate change over the industrial period, which is dominated by uncertainty in forcing by aerosols. For uncertainty in climate sensitivity to be meaningfully reduced and bounded, the present uncertainty in aerosol radiative forcing must be reduced at least three-fold, to less than 0.5 W/m^2.

temperature over the industrial period. Both approaches require knowledge of radiative forcing of Earth's climate over the industrial period.

In addition to radiative forcing by increased CO_2, human activities have resulted in other radiative influences on climate that are of comparable magnitude and that have occurred over roughly the same time period. These include forcing by increased atmospheric mixing ratios of other long-lived GHGs, principally methane, nitrous oxide, chlorofluorocarbons, and tropospheric ozone (O_3), and a decrease in greenhouse forcing by stratospheric O_3. Anthropogenic aerosols cause the other major forcing. Atmospheric aerosol particles scatter and absorb shortwave (solar) radiation (direct effects) and modify the reflective properties and persistence of clouds (indirect effects).[6] Principal aerosol forcing components are sulfates (SO_4^{2-}; cooling), organic carbon (cooling), and black carbon (warming), all caused largely by emissions associated with fossil fuel combustion, and organics and black carbon from biomass combustion (cooling and warming, respectively). Mineral dust from disturbed soils both scatters shortwave radiation (cooling) and absorbs it (warming, if over brighter surfaces). The forcings by these aerosol components are estimated based on loadings calculated by chemical transport models (so-called "forward" calculations[7]) evaluated by comparison with observation. There are likely to have been other, relatively minor, secular changes in radiative flux components caused by aviation-induced contrails and cirrus, changes in surface reflectivity caused by land-use changes, and changes in solar luminosity. These influences are summarized in Figure 1, which shows the estimates by the IPCC working group on radiative forcing[1] of these radiative forcings over the industrial period and of the associated uncertainties. The thesis of this article is that these uncertainties in forcing are too great to be useful for inferring climate sensitivity empirically from observed change in global mean

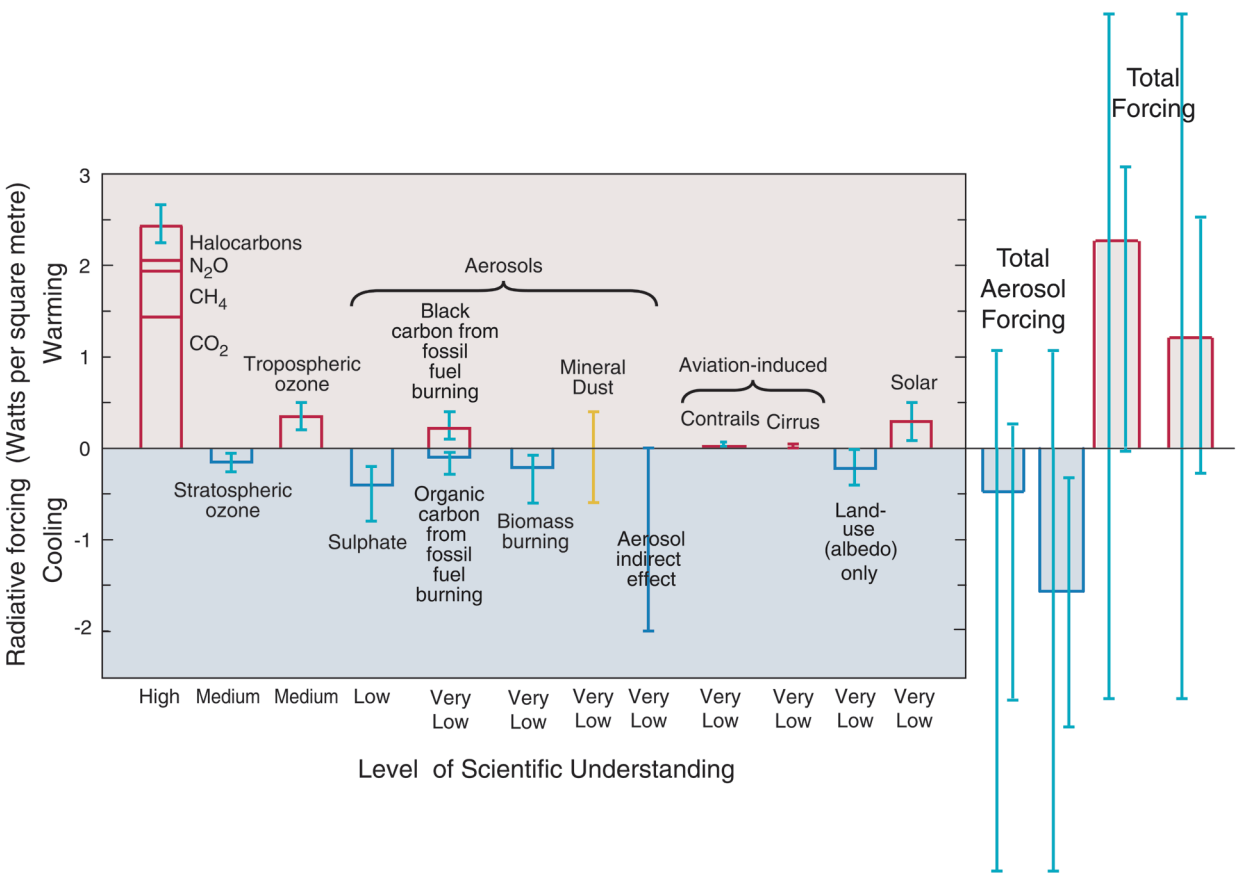

Figure 1. Global mean radiative forcing of climate change for 2000 relative to 1750 as given by the IPCC[1] shown in the left portion of the figure. Positive bars denote warming forcings; negative bars denote cooling forcings; I-beams denote estimated uncertainties. For mineral dust and the aerosol indirect effect, no estimates of the forcing were given, only uncertainty ranges. Level of scientific understanding represents the subjective judgment of the IPCC working group on radiative forcing of the reliability of the forcing estimate. Bars and I-beams at right denote estimates of total aerosol forcing, total forcing, and associated uncertainties. First bar denotes total aerosol forcing evaluated as algebraic sum of IPCC aerosol forcings, with mineral dust and aerosol indirect forcings taken as 0; for second bar, these forcings are taken as the midpoints of the IPCC uncertainty ranges. Third and fourth bars denote total forcing evaluated in the same way, again with mineral dust and aerosol indirect forcings taken as 0 and as the midpoints of the IPCC uncertainty ranges, respectively. For each bar, two uncertainty estimates are provided. Upper and lower limits of the first (larger) uncertainty range are calculated as algebraic sum of upper and lower limits, respectively, of the uncertainties of the several forcings. Upper and lower limits of the second (smaller) uncertainty range are calculated as the square root of the sum of the squares, respectively, of the of the upper and lower uncertainty ranges relative to the estimated forcings denoted by the bars.

temperature over the industrial period or for evaluating performance of climate models over this period by such comparisons.

RADIATIVE FORCING OF CLIMATE CHANGE

The radiative forcing concept is rooted in studies with general circulation models (GCMs) that have examined climate change in response to different kinds of radiative forcing. A consistent finding of such studies is that the change in global mean temperature per forcing, that is the climate sensitivity, is, to a good approximation, independent of the nature of the forcing, for example, forcing because of changes in CO_2 mixing ratios, mixing ratios of other GHGs,[8] aerosol direct forcing,[9] or the solar constant,[10] and independent as well of the geographical distribution of the forcing.[9] These model-based studies are the basis of the climate-sensitivity hypothesis.

An immediate consequence of the forcing-response paradigm is that forcings are additive. This hypothesis provides a path forward to calculating radiative forcing over the industrial period by adding the forcings, as shown in the bars at the right of Figure 1. The first bar gives the algebraic sum of the several aerosol forcings indicated by bars in the IPCC figure. The resulting forcing, -0.46 W/m^2, is small compared with the forcing due to the well-mixed GHGs, 2.42 W/m^2, suggesting slight aerosol influence over the industrial period. However, a problem arises in estimating this forcing because the IPCC working group declined to give a best estimate for forcing by mineral dust and for aerosol indirect forcing. The absence of a number implicitly sets its estimate to 0. In the case of the dust, the uncertainty range is roughly symmetric about 0, so taking the best estimate of this forcing as 0 does not greatly affect the value of the total aerosol forcing. However, for the indirect aerosol forcing, 0 is at one end of a large uncertainty range, so setting this (negative) forcing to 0 has the effect of biasing the total to far too great a positive value.

The IPCC working group on radiative forcing[1] estimated the range of the aerosol indirect forcing over the industrial period between 0 and -2 W/m^2. Table 6.6 of the radiative forcing chapter of the IPCC report lists some 17 independent estimates ranging from -0.3 to -1.85 W/m^2, and other even greater (negative) estimates have been published; the working group extended the range upward to 0 to account for possible absorption by black particles in clouds. In any event, for the purpose of estimating total aerosol forcing over the industrial period, it seems certain that 0 leads to a biased estimate. Using the midpoints of the uncertainty ranges for both the mineral dust forcing and the aerosol indirect forcing results in the total aerosol forcing (second bar at the right of Figure 1) being -1.56 W/m^2, almost two-thirds of the forcing by long-lived GHGs. Also shown are estimates of the total radiative forcing calculated in the same way, with mineral dust and aerosol indirect forcings taken as 0 (bar 3) and as the midpoints of the uncertainty ranges (bar 4). The latter estimate, 1.21 W/m^2, is roughly half the total forcing obtained if these two forcings are taken as 0. It should be evident that in any empirical inference of climate sensitivity or in any evaluation of performance of climate models over the industrial period, the result will be quite sensitive to the choice of aerosol forcing.

Also shown for each estimate of total aerosol forcing and total forcing are two estimates of the associated uncertainty. For the left I-beam of each pair, the upper and lower limits are calculated as the algebraic sums of the upper and lower limits, respectively, of the uncertainties associated with each of the individual forcings. For the right I-beam, the limits are calculated as the square root of the sums of the squares of the differences between the limits of the respective uncertainty ranges and the estimated forcings (RSS method). The algebraic sum, which leads to an upper limit uncertainty in the total forcing, would be appropriate for positively correlated errors, as might occur, for example, if atmospheric residence times of aerosol particles used in estimating both the direct and indirect aerosol forcing were systematically greater than the central values on which present forcing estimates rest. The RSS method would be appropriate for uncorrelated errors. An alternative means of propagating these uncertainties was presented by Boucher and Haywood,[11] who used a Monte Carlo approach for several assumed shapes of the probability distribution functions (PDFs) of the uncertainties to obtain PDFs of the overall uncertainty range. The peaks of the PDFs for the total forcing were at $\sim$1 W/m^2, with the bulk of the PDFs between 0 and 2 W/m^2 and with some probability (3–25%, depending on assumption in propagating the uncertainty) that the total forcing is negative. By any reckoning, the uncertainty in the total radiative forcing of climate change over the industrial period, which is dominated by the uncertainty in total aerosol forcing ($\sim \pm 1.35$ W/m^2 by the RSS method) is quite large relative to the forcing by greenhouse gases.

It is useful to examine the reasons for the large uncertainties associated with the several forcings. Here, the comparison between the well-mixed GHGs and tropospheric O_3 is instructive. The relative uncertainty for GHGs is much less than that for tropospheric O_3, yet the absolute uncertainties are comparable. The present and preindustrial atmospheric mixing ratios of the well-mixed GHGs are well established, from contemporary measurements at a relatively small number of locations and preindustrial measurements from ice cores, and the uncertainty in forcing is thus caused largely by issues of radiation transfer, as exemplified by differing estimates of forcing by CO_2 (see Appendix).

For tropospheric O_3, the uncertainty in forcing is dominated by uncertainty in the increase in mixing ratio of this relatively short-lived and heterogeneously distributed

atmospheric species above its preindustrial value. Similar considerations apply to the aerosol forcing. The large relative uncertainty in the direct forcing by SO_4^{2-} is caused mainly by issues of atmospheric chemistry—how much is formed, how long the material resides in the atmosphere—and aerosol microphysics—the size distribution and its response to relative humidity.[12] Relatively little uncertainty is caused by uncertainty in the treatment of the radiative effects of this aerosol, provided the loading and size distribution are specified.[13]

Climate forcing is quite sensitive to incremental aerosol loading. An increase in aerosol scattering optical depth of 0.03 in cloud-free areas of the planet, an amount close to the limit of detection of well-calibrated instruments,[14,15] less than the natural aerosol optical depth in pristine regions of the southern hemisphere,[14] and well less than day-to-day fluctuations in rural, mid-continental North America,[15] gives rise[6,13] to a radiative forcing of ~ -1 W/m^2. The indirect aerosol effect, which is caused by an increase in multiple scattering in and reflectivity of clouds of intermediate optical depth caused by the presence of enhanced concentrations of aerosol particles that serve as nuclei for cloud drop formation,[16] is quite sensitive to cloud drop number concentration; an increase in 30% of the number concentration of cloud droplets in marine stratus clouds is estimated[6] to yield a global average radiative forcing of ~ -1 W/m^2.

APPROACHES TO DETERMINING CLIMATE SENSITIVITY

As noted previously, two independent approaches might be taken to determine sensitivity of global mean temperature to a radiative forcing, studies with climate models and empirical inference. In principle, if climate models were accurate and complete, they might be used with confidence to predict the response of future climate to future forcing. In practice, these models embody numerous assumptions, parameterizations, and approximations of the variables and phenomena being represented—water vapor, clouds, precipitation, snow and ice, radiation, transport of heat and water on all scales—the list goes on. The resolution of models is limited, typically, at present to ~300 km, with the necessity to parameterize phenomena occurring at smaller scales. Much model evaluation is carried out by examination of the ability of models to simulate the present climate.[17] However, even models that reproduce many aspects of the present climate rather well still can exhibit widely differing climate sensitivities. The IPCC 2001 survey[18] of 15 current atmospheric GCMs that are coupled to mixed-layer upper ocean models reported a range of sensitivities to doubled CO_2, $\Delta T_{\times 2}$, of 2–5.1 K; average 3.5 K; standard deviation 0.92 K. Modeled climate sensitivity is highly dependent on parameterizations; for example, a change in cloud parameterization in the UK Meteorological Office model changed modeled climate sensitivity to doubled CO_2 by a factor of 2.8, from 1.9 to 5.4 K.[19] The wide range in modeled climate sensitivities and the sensitivity to parameterization suggest that confidence in the ability of climate models to predict climate change in response to future forcings cannot be based only on their ability to simulate present climate but rather is to be gained from their ability to reproduce climate change over the industrial period.

Recent studies using coupled atmosphere-ocean models have included aerosol forcing in an effort to simulate transient climate response over the industrial period, some results of which are shown in Figures 2 and 3 and summarized in Table 1. In such studies, it is necessary to assume a time profile (and spatial distribution) of the forcings to be represented in the model. The model is then run subject to these time-dependent forcings and compared with a base case for which no forcing is applied. Because of internal variability within such models, it is necessary to carry out some averaging. For studies with constant forcing differences, this variability can be accounted for by running the model for sufficient time to obtain a stable average, but this cannot be done for runs with transient forcings, so internal variability is estimated by running ensembles of several transient calculations. Model performance is evaluated by comparison with the temperature record observed over the industrial period. With inclusion of aerosol forcing, each of the models fairly accurately reproduces this record; the modeled temperature increase over the period 1900–1990, for example (Table 1), is similar to the observed[20] increase over this period, 0.48 ± 0.17 K (2σ). Without exception, the investigators assert that inclusion of aerosol forcing improves the comparison of their model simulations to the observed temperature trend. However, it would seem that confidence in such agreement and, in turn, in the correspondence between the temperature sensitivity employed in any given model and the actual sensitivity of Earth's climate system, must be tempered by the uncertainty in total forcing at present (and a fortiori as a function of time over the industrial period) and by the differing sensitivities of the several models (by up to a factor of 1.7) and aerosol forcings (by up to a factor of 1.8) employed in these studies. The intermodel spread in modeled temperature trend over the period 1900–1990 expressed as a fractional standard deviation is much less than the corresponding spread in either sensitivity or aerosol forcing (Table 1). Certainly a major contributor, if not the greatest contributor, to the inability of comparisons between modeled and observed temperature trends to narrow down the range of model sensitivity is the uncertainty in radiative forcing over the industrial period, which is dominated by uncertainty in radiative forcing by aerosols.

Considerations such as the foregoing would seem to call into question confidence that can be placed in

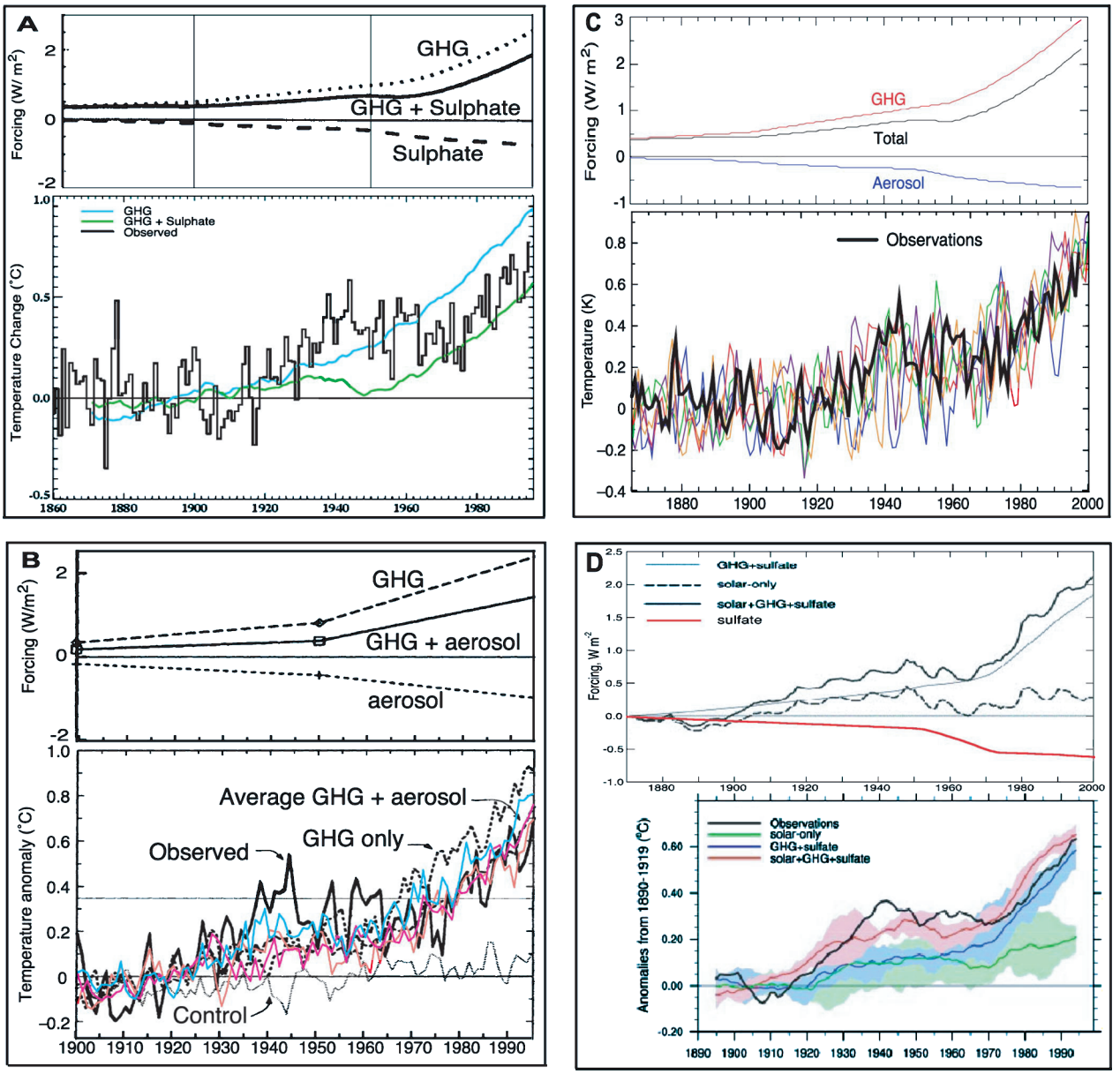

Figure 2. Global mean forcing and temperature anomaly in four climate models over the past 100–140 yr. (a) UK Meteorological Office model, 1995;[25] (b) Canadian Climate Model, 2000;[26] (c) Geophysical Fluid Dynamics Laboratory model, 2000;[27] (d) National Center for Atmospheric Research model, 2000.[28] Multiple traces for forcings denote different combinations of imposed forcings. Multiple traces for temperature change in (b) and (c) and envelopes in (d) denote results from ensembles of multiple model runs with identical forcings and similar but slightly different initial conditions. Observed temperature anomaly, bold black. For model sensitivities, aerosol forcings, and investigator characterization of model performance, see Table 1.

statements such as the following from the IPCC 2001 assessment of climate change:[21]

- "Simulations that include estimates of natural and anthropogenic forcing reproduce the observed large-scale changes in surface temperature over the 20th century."
- "Most model estimates that take into account both greenhouse gases and sulphate aerosols are consistent with observations over this period."
- "The large-scale consistency between models and observations can be used to provide an independent check on projected warming rates over the next few decades under a given emissions scenario."
- "Detection and attribution studies comparing model-simulated changes with the observed record can now take into account uncertainty in the magnitude of modeled response to external forcing, in particular that be due to uncertainty in climate sensitivity."

The alternative approach to determining model sensitivity is empirical. In essence, climate sensitivity would be evaluated as $\lambda = \Delta T/F$, where ΔT is the observed global-average temperature change over some period of record and F is the

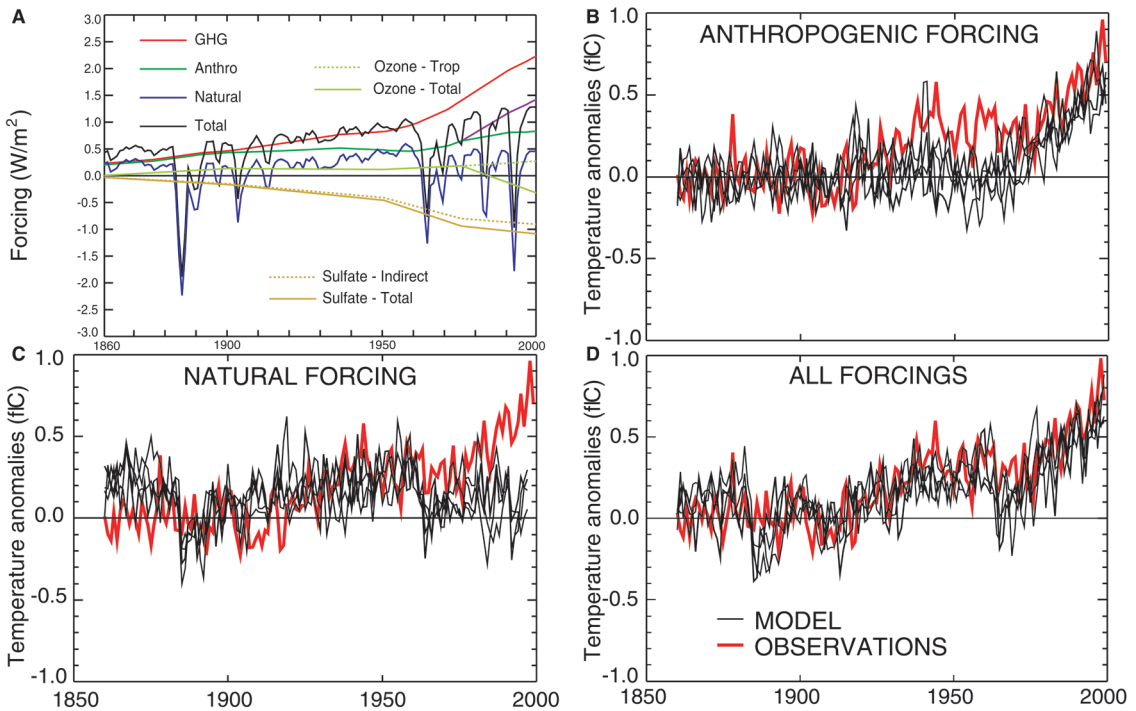

Figure 3. Global mean forcing and temperature anomaly in the UK Meteorological Office 2000 model[29,30] for the period 1860–1998. Results of four model runs are shown for each indicated combination of forcings together with observed temperature anomaly (red). For model sensitivity, aerosol forcing, and investigator characterization of model performance, see Table 1.

forcing over that period, corrected[18] to account for loss of heat from the surface to the deep ocean resulting from the fact that the system is not at equilibrium, by decreasing the forcing by dH_o/dt, the rate of increase of heat energy H_o stored by the deep ocean. The period of record might be the period from 1860 to the present, for which fairly reliable instrumental records are available, or might be some shorter period that is sufficiently long to allow coupling of the atmosphere to the ocean mixed layer (~2 years) and to average over short-term variability due to El Niño cycles and the like (at least a decade). By analyzing measurements of ocean temperature as a function of location, depth, and time, Levitus et al.[22] determined the increase in heat content of the upper 3000 m of the global oceans over the period 1951–1996. The heat flow to the deep ocean dH_o/dt given by the slope of a linear fit to the time-dependent heat storage for the world ocean is 0.23 ± 0.055 W/m^2, where the uncertainty is the standard error in the slope for measurements assumed to be independent at 5-yr intervals. For purposes of empirical evaluation of the equilibrium climate sensitivity, a negative forcing of this magnitude and uncertainty must be added to the other forcings and uncertainties indicated in Figure 1. Like the model-based approach, this empirical approach to determining climate sensitivity is limited by considerations of uncertainty, again mainly uncertainty in forcing. For example, Gregory et al.[23] found that the uncertainty in forcing was consonant with arbitrarily large sensitivity as (negative) aerosol forcing was increased such that the total forcing approached 0.

ESTIMATING THE REQUISITE UNCERTAINTY IN AEROSOL FORCING

For climate sensitivity to be evaluated empirically, the fractional uncertainty in λ can be estimated to first-order as

$$\frac{\delta\lambda}{\lambda} = \sqrt{\left(\frac{\delta\Delta T}{\Delta T}\right)^2 + \left(\frac{\delta F}{F}\right)^2} \qquad (1)$$

where δF and $\delta\Delta T$ denote the uncertainties in F and ΔT, respectively. A reasonable initial target uncertainty might be $\delta\lambda/\lambda = 30\%$; for example:

$$\Delta T_{2\times CO_2} = (3 \pm 1)\ \mathrm{K} \qquad (2)$$

Setting such a target allows specification of required uncertainties in temperature anomaly and forcing. For example, the required uncertainty in λ might be met for $\delta\Delta T/\Delta T \approx \delta F/F \approx 20\%$. An uncertainty of such magnitude is arguably already at hand for the temperature difference over the instrumental period (1861–2000), for which Folland et al.[24] estimate a fractional uncertainty (2σ) of 26%.

The requirement of a specific uncertainty in total forcing over such a period, for example, 20%, allows the required uncertainty in aerosol forcing to be estimated. Here,

Table 1. Performance of climate models that include aerosol forcing in transient simulations.

Model, Year	Sensitivity[a] $\Delta T_2 \times$ K	1990 Aerosol Forcing W/m^2	Nature of Aerosol Forcing	Modeled ΔT^b K	Investigator Characterization of Model Performance Including Aerosol Forcing
UK Met Office[25] 1995	2.5	−0.6	Surface albedo enhancement in lieu of direct forcing by SO_4^{2-} aerosol, geographically distributed per chemical transport model calculations[31] and temporally scaled per emissions[32,33]	0.48	Inclusion of sulfate aerosol forcing improves the simulation of global mean temperature over the last few decades
Canadian Climate Model[26] 2000	3.5	−1	Surface albedo enhancement geographically distributed per chemical transport model calculations[31] with linear temporal ramp from 0 in 1850	0.56	Observed global mean temperature changes and those simulated for GHG + aerosol forcing show reasonable agreement
GFDL[27] 2000	3.4	−0.62	Similar to UK Met Office (1995)[25]	0.46	The surface temperature time series from the five GHG-plus-sulfate integrations show an increase over the last century, which is broadly consistent with the observations
UK Met Office[28,29] 2000	3.35	−1.1	Fully interactive sulfur cycle that represents the emission, transport, oxidation, and removal of sulfur species and direct and indirect effects	0.38[c]	The ALL ensemble captures the main features of global mean temperature changes observed since 1860
NCAR[30] 2003	2.18	−0.6	SO_4^{2-} aerosol direct effects only; SO_4^{2-} loadings from runs with NCAR's Climate System Model with interactive sulfur chemistry and direct input of SO_2 emissions	0.5	The time series from GHG + sulfates + solar shows reasonable agreement with the observations
Range	1.32	0.5		0.18 0.1[d]	
Fractional standard deviation, %	20	31		14 9[d]	

[a]As stated by the investigators; [b]Modeled change in global mean temperature between 1900 and 1990 for greenhouse gas and aerosol forcing; [c]Total anthropogenic forcing; also includes stratospheric O_3 forcing −0.4 W/m^2; [d]Omitting UK Met Office (2000), which also included stratospheric O_3 forcing.

consideration is restricted just to forcing by GHGs F_g and aerosols F_a, setting aside other contributions to forcing and uncertainty. For this estimate, total GHG forcing (well-mixed gases plus stratospheric and tropospheric O_3) is taken equal to 2.6 W/m^2 with an uncertainty of ±10% as given by the IPCC (Figure 1). The total forcing is the algebraic sum $F = F_g + F_a$, but because aerosol forcing is negative the relative uncertainty in the total forcing, evaluated as

$$\delta F/F = (F_g^2 + F_a^2)^{1/2}/F \tag{3}$$

will be greater than that in either term; that is,

$$\delta F_g/F_g \text{ and } \delta F_a/F_a \tag{4}$$

and if the two terms are comparable in magnitude, much greater. The requirement that uncertainty in aerosol plus greenhouse gas forcing not exceed 20% permits specification of an upper bound on the uncertainty in aerosol forcing; this requisite uncertainty is shown in Figure 4. Here, the left ordinate gives the maximum uncertainty in aerosol forcing consistent with the uncertainty in the total forcing not exceeding 20% as a function of (negative) aerosol forcing on the abscissa. No matter what the magnitude of the aerosol forcing, the requirement that the uncertainty in total forcing not exceed 20% means that the uncertainty in aerosol forcing cannot exceed ~±0.45 W/m^2; the requirement on the aerosol uncertainty becomes increasingly stringent as the magnitude of (negative) aerosol forcing increases; for aerosol forcing ~ −1.2 W/m^2, which is still less than half that of GHG forcing and well within the estimated range, the requisite uncertainty is ±0.11 W/m^2. These required uncertainties can be compared with the present conservatively estimated value of ±1.35 W/m^2 (Figure 1). The right axis gives this maximum uncertainty as a fractional uncertainty on a nonlinear scale. For small aerosol forcing, the fractional uncertainty in this forcing can be quite high and still meet the requirement that the uncertainty in total forcing not exceed 20%, but this fractional uncertainty decreases strongly as (negative) aerosol forcing increases, becoming the same as that currently ascribed to GHG forcing, that is, ~10%, at an aerosol forcing of ~ −1.2 W/m^2. Essentially the same uncertainty in aerosol forcing would be required, as a function of secular time, as input to GCM

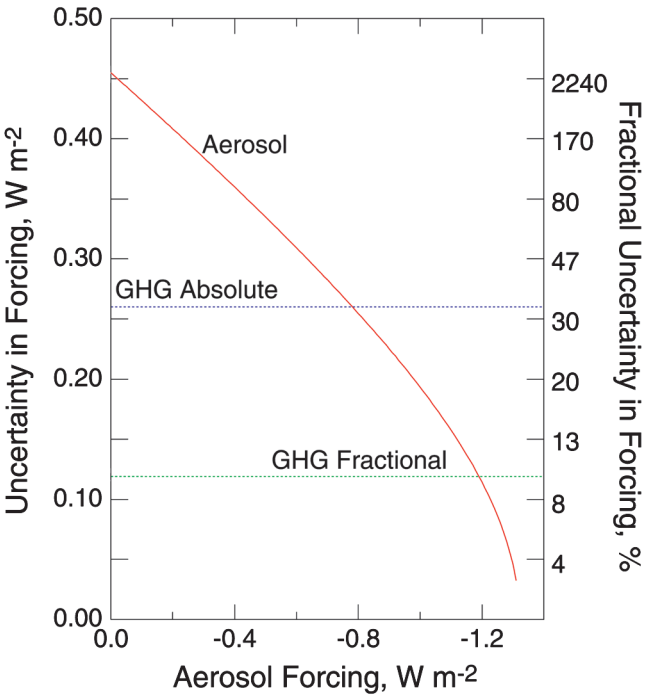

Figure 4. Absolute (left axis) and relative (right axis) uncertainty in radiative forcing of climate change by anthropogenic aerosols (red) required to achieve uncertainty of 20% in total radiative forcing over the industrial period, and in turn, uncertainty of 30% in climate sensitivity as a function of (negative) aerosol forcing. Blue and green lines denote, for comparison, the IPCC estimate[1] of present uncertainty in GHG forcing.

calculations of the temperature anomaly trend over the industrial period.

Reducing the uncertainty in aerosol forcing to such an extent would represent an enormous challenge to the atmospheric research community. Nonetheless, the need for reducing the uncertainty from its present estimated value by at least a factor of 3 and perhaps a factor of 10 or more seems inescapable if the uncertainty in climate sensitivity is to be reduced to an extent where it becomes useful for formulating policy to deal with global change. If this challenge is not met, it is likely that in another 20 years it will still not be possible to specify the climate sensitivity with uncertainty range appreciably narrower than it is at present.

The uncertainty in aerosol forcing implies that Earth's climate sensitivity may be substantially greater, or substantially less, than would be calculated based on observed temperature rise over the industrial period and estimates of aerosol forcing shown in Figure 1. If the magnitude of (negative) aerosol forcing is at the high end of the uncertainty range, then aerosol forcing is offsetting a substantial fraction of GHG forcing, and Earth's climate sensitivity is at the high end of the range suggested by IPCC[2,3] or perhaps is greater. To whatever extent aerosol forcing is offsetting GHG forcing, the greatly disparate atmospheric residence times of GHGs (decades to centuries) and of tropospheric aerosols (days) implies that any reliance on aerosol forcing to offset continuously increasing GHG forcing in the future would require an indefinite commitment to continuously increasing aerosol concentrations. These considerations have inevitable implications on formulation of policy regarding mitigation of climate change or adaptation to greenhouse warming.

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APPENDIX

Unit of Measure for Climate Sensitivity

Any examination of uncertainty in climate sensitivity must consider the unit of measure. Traditionally, the sensitivity of climate change has been expressed as a quasi-equilibrium change in global mean temperature that would result from a doubling of atmospheric mixing ratio of CO_2, $\Delta T_{2\times}$. The temperature sensitivity, a change in temperature due to a change in a radiative flux component, has dimension temperature/(power per area), for which the systematic unit is K/(W/m^2); this quantity is commonly denoted by the symbol λ. Such a systematic unit for sensitivity is to be preferred because it is insensitive to uncertainty in the forcing caused by doubling of CO_2. The conversion is given by

$$\lambda = \Delta T_{2\times}/F_{2\times}$$

where $F_{2\times}$ is the forcing of doubled CO_2 relative to the unperturbed state, ~4 W/m^2. The use of $\Delta T_{2\times}$ as a unit of climate sensitivity implies that $F_{2\times}$ is known with small uncertainty and independent of the CO_2 mixing ratio of the unperturbed state; more importantly, comparison of sensitivities of different models assumes that all investigators obtain the same forcing for such a doubling.

The forcing associated with doubling of CO_2, $F_{2\times}$, was examined in an intercomparison of 15 atmospheric general circulation models reported by Cess et al.[34] In that study, the base case was specified as 330 ppm and the perturbed case was taken as 660 ppm, eliminating any contribution to variation that might have resulted from other choices. Likewise, that intercomparison specified that the forcing was to be calculated as a net flux difference at a pressure altitude of 200 hPa, again potentially reducing the variation in the stated forcing associated with doubling of CO_2. Nonetheless, the variation among the several models was substantial, with range 34% and fractional standard deviation 10.2%. Cess et al. noted that if the 15 models all exhibited equal climate sensitivity, λ = 1 K/(W/m^2), this variation in forcing per CO_2 doubling would lead to change in equilibrium temperature for doubled CO_2 ranging from 3.4 to 4.7 K and pointed out that this range is nearly half the often quoted range of uncertainty in $\Delta T_{2\times}$ of 1.5–4.5 K. More recently, the 2001 assessment report of the IPCC[1] revised the forcing associated with a doubling of CO_2 from 4.37 W/m^2 used in previous reports to 3.7 W/m^2, a decrease of 15%, with implicit resultant changes in climate sensitivities reported in terms of doubled CO_2. The uncertainty associated with the use of $\Delta T_{2\times}$ as a measure of climate sensitivity is both substantial and wholly unnecessary. Nonetheless, because of the widespread prior use of $\Delta T_{2\times}$ as a measure of sensitivity, it seems useful to retain this measure, provided it is viewed as subsidiary and is uniquely and unambiguously related to the sensitivity λ in systematic units K/(W/m^2).

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