Connections between aerosol forcing (and its uncertainty) and climate sensitivity
Uncertainties in aerosol forcings. Global-mean anthropogenic aerosol forcing over the industrial era (left axis) as estimated by forward (A to F) and inverse (G to L) calculations and as used in applications (M to Q) (20). Circles with error bars are central values and 95% confidence limits. Bare error bars are stated range. Squares represent specific forcing calculations using alternative formulations within the same study. Right axis: Total forcing over the industrial era using the approximation that nonaerosol forcings are 2.7 W m$^{-2}$ (3, 4).

Note: Applications use forcings derived from inverse models so are not independent

Strong present-day aerosol cooling implies a hot future

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Atmospheric aerosols counteract the warming effects of anthropogenic greenhouse gases by an uncertain, but potentially large, amount. This in turn leads to large uncertainties in the sensitivity of climate to human perturbations, and therefore also in carbon cycle feedbacks and projections of climate change. In the future, aerosol cooling is expected to decline relative to greenhouse gas forcing, because of the aerosols' much shorter lifetime and the pursuit of a cleaner atmosphere. Strong aerosol cooling in the past and present would then imply that future global warming may proceed at or even above the upper extreme of the range projected by the Intergovernmental Panel on Climate Change.


Twentieth century climate model response and climate sensitivity

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[1] Climate forcing and climate sensitivity are two key factors in understanding Earth’s climate. There is considerable interest in decreasing our uncertainty in climate sensitivity. This study explores the role of these two factors in climate simulations of the 20th century. It is found that the total anthropogenic forcing for a wide range of climate models differs by a factor of two and that the total forcing is inversely correlated to climate sensitivity. Much of the uncertainty in total anthropogenic forcing derives from a threefold range of uncertainty in the aerosol forcing used in the simulations. Citation: Kiehl, J. T. (2007), Twentieth century climate model response and climate sensitivity, Geophys. Res. Lett., 34, L22710, doi:10.1029/2007GL031383.

[4] One curious aspect of this result known [Houghton et al., 2001] that agree in simulating the anomaly in climate sensitivity from most global climate models used for a range in climate sensitivity from 1.5 to 4.5°C for a dc this range commonly varies by at least a factor of two in eqn [5] The question is: if climate model 2 to 3 in their climate sensitivity, how the global temperature record with accuracy. Kerr [2007] and S. E. Schwi climate change, too may a picture.

Key messages

• Strong (negative) aerosol forcing implies a higher sensitivity of the global mean temperature to CO$_2$ increases (the “equilibrium climate sensitivity”, or ECS)

• Models achieve the correct 20$^{th}$ century temperature trends by compensation between two highly uncertain components: ECS and aerosol radiative forcing, both of which are set by cloud responses
Box 1 | Climate sensitivity

The problem of predicting global climate change can be symbolically represented by a simple heat balance equation:

$$ \frac{d(\Delta T)}{dt} = \Delta Q - \lambda \Delta T $$

Here $\Delta T$ is the global mean temperature change arising from a change in radiative forcing $\Delta Q$. $\Delta Q$ represents the total climate forcing (in W m$^{-2}$) due to changes in natural factors (such as volcanoes and solar variability), as well as human-induced changes in the concentrations of greenhouse gases and aerosols. The long-term equilibrium response to a radiative forcing (such as doubling of CO$_2$) is given by the parameter $\lambda$ as follows: $\Delta T_{2\times CO_2} = \Delta Q_{2\times CO_2} / \lambda$, where $\Delta Q_{2\times CO_2} = 3.7$ W m$^{-2}$. $\lambda$ itself depends on many climate feedback processes, such as those arising from changes in water vapour, snow cover and clouds. The left-hand side of this equation is a heat storage term which determines how quickly the climate system approaches this equilibrium state. The heat capacity $c$ can be estimated from observations of ocean heat uptake$^{24}$ and recent warming trends$^{13}$ as $1.1 \pm 0.5$ GJ m$^{-2}$ K$^{-1}$. However, there is a wide range in projections of future climate change primarily because of uncertainties in both $\lambda$ and the future $\Delta Q$. 
Kiehl (right) shows total forcing (20th century), which is the sum of the various radiatively-active constituent changes:
\[ \approx 2.4 \pm 0.3 \text{ Wm}^{-2} \text{ from GHGs + aerosol forcing} \]

Figure 1 | Climate sensitivity required to explain the observed 1940–2000 warming as a function of the strength of aerosol radiative cooling. The solid line represents results using the central estimate of heat capacity \((1.1 \pm 0.5 \text{ GJ m}^{-2} \text{ K}^{-1})\) from Levitus et al., and the dashed (dot-dashed) lines represent the higher (lower) limit of this heat capacity. More details of the model are given in Box 3.

Figure 1. Total Anthropogenic Forcing (Wm\(^{-2}\)) versus equilibrium climate sensitivity (°C) from nine coupled climate models and two energy balance models that were used to simulate the climate of the 20th century. Solid line is theoretical relationship from equation (4). Dashed lines are from assuming a ±0.2 Wm\(^{-2}\) uncertainty in ocean energy storage in equation (4).
IPCC “Radiative Forcing” Diagram

Understand physical bases for, and model representation of, this entry.

IPCC 5th Assessment Report (AR5), Summary for Policymakers, Fig. SPM.5, 2013.
Box 2 | Future aerosol scenarios

The SRES emissions scenarios used in the IPCC-TAR all suggest that aerosol emission by the middle of this century will be near or below present levels. Because aerosols are very short-lived in the atmosphere—lifetimes of days compared with decades for the greenhouse gases—they do not accumulate and the burden is almost proportional to the emissions. Consequently, as we clean up our vehicles and smokestacks to provide cleaner air and improve air quality, the aerosol loading of the atmosphere will decrease. Even population growth and increasing industrialization in the developing countries will do little to change this outcome. We are already in the process of revising downward our projections of aerosol emissions from China and other developing countries, as they are introducing cleaner technology faster than had been anticipated a decade or so ago. Because of the rapidly growing knowledge of the very serious health effects of aerosols we expect that regulatory efforts will act to reduce aerosol emissions even more rapidly than anticipated when the SRES scenarios were developed.

Box 2 Figure | Historical CO₂ and SO₂ emissions from 1850–2000, followed by projected values to the year 2100 from the SRES A2 scenario.
Figure 2 | Temperature change simulated by the simple model for the period 1850 to 2100. Two extreme cases are shown: strong present-day aerosol cooling consistent with ‘forward’ studies of aerosol effects on climate but with a climate sensitivity not ruled out by observations (red line, $Q_{\text{aeros}} = -1.7 \text{ W m}^{-2}$), and the case of no aerosol cooling effect (blue line). The shading and the yellow line represent the range and central projection given in IPCC-TAR, based on the same scenario used in these calculations (scenario A2, from ref. 25).
Box 3 | Simple climate–carbon cycle model

We use a zero-dimensional climate–carbon cycle model\textsuperscript{29}, which updates the global temperature using the equation in Box 1 and accounts for potentially large positive carbon cycle feedbacks by updating CO\textsubscript{2} interactively on the basis of the emissions scenario. It uses a simple fit to the ocean and land uptake of CO\textsubscript{2} derived from the Hadley Centre’s climate–carbon cycle GCM\textsuperscript{9}, but with alternative sets of possible land sensitivity parameters chosen to fit the observed CO\textsubscript{2} rise. The land carbon cycle responses produced by this simple fit therefore span the range simulated by other potentially realistic models. The size of the climate–carbon cycle feedback depends critically on the opposing effects of CO\textsubscript{2}-fertilization on plant growth, and enhanced soil decomposition as the climate warms. The latter is dependent on the degree of climate warming, as well as the sensitivity of soil respiration to temperature\textsuperscript{18,29}.

The major anthropogenic forcings are considered: CO\textsubscript{2}, other well-mixed GHGs, and sulphate aerosols. The radiative forcing from CO\textsubscript{2} and other well-mixed GHGs are derived from well-known formulae\textsuperscript{1}. The radiative forcing from sulphate aerosols is assumed to be proportional to global mean sulphate loading, which in turn is assumed to be proportional to SO\textsubscript{2} emissions. To avoid undue influence of other forcing factors (in particular natural forcing from solar and volcanic sources) we consider just the portion of the historical record that is dominated by anthropogenic influence—namely 1940 to present\textsuperscript{30}. 
• Uncertainty in aerosol forcing means that within-scenario warming uncertainty is larger than that between scenarios.

• Strong ECS and therefore stronger warming also leads to a smaller uptake of CO$_2$ and further warming (positive carbon cycle feedback).

Figure 3 | Modelled temperature change and CO$_2$ increase by 2100 under different development scenarios. a, Temperature rise by the year 2100 for the various SRES scenarios$^{35}$ as a function of present-day aerosol cooling. The horizontal green bar indicates the threshold of ‘dangerous anthropogenic interferences’ in the sense of the United Nations Framework Convention on Climate Change, using an estimate of 1.5–2°C for this value, based on arguments by different groups$^{26,27}$. At all but the very lowest climate sensitivities this level will be exceeded unless GHG emissions are reduced below those in the SRES B1 and B2 scenarios (from ref. 25). The pink bar indicates the temperature change between ice ages and interglacials$^{38}$. b, Same as a but for atmospheric CO$_2$ by 2100. The shaded areas represent the IPCC-TAR range across models and scenarios.
Figure 4 | Strength of climate-carbon cycle feedback as a function of climate sensitivity. The feedback factor is defined as the CO₂ concentration rise projected between 1860 and 2100, divided by the CO₂ rise predicted in the absence of climate effects on the carbon cycle. Results are shown for the A2 scenario as a function of climate sensitivity/present-day aerosol cooling, for various sensitivities of soil respiration to temperature. The shaded area represents the range of feedback factor estimated from the IPCC-TAR range of CO₂ concentrations relative to the standard A2 concentration scenario. \( q_{10} \) values of 1, 2, 3 and 4 correspond to \( C_{0.5} \) values of 295, 485, 676, 866 p.p.m., respectively, in order for the model to recreate the observed CO₂ record.
uncertainty has decreased. Here we show, based on data from the two reports, that this evolution towards lower uncertainty can be expected to continue into the future. Because it is easier to reduce air pollution than carbon dioxide emissions and because of the long lifetime of carbon dioxide, the less uncertain carbon dioxide forcing is expected to become increasingly dominant. Using a statistical model,