LETTERS

An Observationally Based Estimate of the Climate Sensitivity

J. M. GREGORY

Hadley Centre, Met Office, Bracknell, Berkshire, United Kingdom

R. J. STOUFFER

Geophysical Fluid Dynamics Laboratory, Princeton, New Jersey

S. C. B. RAPER

Climatic Research Unit, University of East Anglia, Norwich, Norfolk, United Kingdom, and Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany

P. A. STOTT AND N. A. RAYNER

Hadley Centre, Met Office, Bracknell, Berkshire, United Kingdom

8 May 2002 and 5 July 2002

ABSTRACT

A probability distribution for values of the effective climate sensitivity, with a lower bound of 1.6 K (5th percentile), is obtained on the basis of the increase in ocean heat content in recent decades from analyses of observed interior-ocean temperature changes, surface temperature changes measured since 1860, and estimates of anthropogenic and natural radiative forcing of the climate system. Radiative forcing is the greatest source of uncertainty in the calculation; the result also depends somewhat on the rate of ocean heat uptake in the late nineteenth century, for which an assumption is needed as there is no observational estimate. Because the method does not use the climate sensitivity simulated by a general circulation model, it provides an independent observationally based constraint on this important parameter of the climate system.

1. Introduction

The equilibrium climate sensitivity is the conventional measure of the equilibrium climate response to radiative forcing resulting from greenhouse gases and other anthropogenic and natural causes. It is defined as the steady-state change in global-average surface temperature due to a doubling of the carbon dioxide concentration, and is estimated to lie between 1.5 and 4.5 K (Cubasch et al. 2001), largely on the basis of experiments with general circulation models (GCMs). This wide range was informally obtained from the model results, and does not correspond to any particular probability limits. Despite considerable improvements in many aspects of the simulation of twentieth-century climate by GCMs, the range has remained essentially unchanged during the last two decades, and is the greatest source of uncertainty in climate change projections for the twenty-first century.

GCMs indicate that the increase in global-average outgoing radiative flux when the climate is perturbed from a steady state is proportional to the global-average surface temperature change ΔT. During time-dependent climate change, the imbalance between the imposed radiative forcing Q and the radiative response $\lambda\Delta T$, λ being a constant, is absorbed by the heat capacity of the system, which resides overwhelmingly in the ocean (Levitus et al. 2001). Hence,

$$F(t) = Q(t) - \lambda\Delta T(t), \tag{1}$$

where t is time and F is the heat flux into the ocean.

Corresponding author address: Dr. J. M. Gregory, Hadley Centre, Met Office, London Road, Bracknell, Berkshire RG12 2SY, United Kingdom.
E-mail: jonathan.gregory@metoffice.com

Equation (1) has often been employed as the basis for energy-balance climate models.

In the unperturbed steady-state climate, $Q = F = 0$ and $\Delta T = 0$. If Q is raised from zero to some positive value, F becomes positive, additional heat is stored in the ocean, and ΔT rises. If Q then remains constant, F returns to zero over time, as the climate approaches a new steady state in which $\Delta T = Q/\lambda$. From its definition, the equilibrium climate sensitivity $\Delta T_{2\times} = Q_{2\times}/\lambda$, where $Q_{2\times}$ is the forcing that results from a doubling of the CO_2 concentration.

Although it is defined in terms of a steady-state climate, the climate sensitivity can be estimated from any climate state. Provided we know F, Q, and ΔT, we can calculate λ from Eq. (1) and, hence, $\Delta T_{2\times}$ (e.g., Cubasch et al. 2001). Some results with coupled atmosphere–ocean GCMs (AOGCMs) suggest that $\Delta T_{2\times}$ (called the ''effective'' climate sensitivity when calculated from an unsteady climate) might not be constant even on the century timescale (Senior and Mitchell 2000), although AOGCM experiments do not give rise to any expectation that it will change rapidly. If $\Delta T_{2\times}$ is not constant, its usefulness for predicting future climate change is of course limited, and an estimate based on recent climate change is the most appropriate one to use. The utility of the climate sensitivity also depends on the response being independent of the nature of the agent causing the radiative forcing.

2. Method

Recent studies aimed at setting constraints on the climate sensitivity have used climate models in which λ can be varied and heat uptake by the ocean is modeled simply (Wigley et al. 1997; Andronova and Schlesinger 2001; Forest et al. 2002). The approach is systematically to adjust the parameters and inputs of the model, comparing the simulated results with observed surface temperature changes. The results give a range for $\Delta T_{2\times}$ that is even wider than 1.5°–4.5°C.

Using a model of ocean heat uptake inevitably involves assumptions about its mechanisms. Estimates of ocean heat uptake can instead be made using the 5-yr running means of observed interior-ocean temperature changes of Levitus et al. (2000). The increase in heat content from 1957 to 1994, the period of best data coverage, is 19.0 ($\pm$ 9.0) $\times$ 10^{22} J. The stated uncertainty, of two standard deviations, relates to measurement and sampling uncertainties. Denoting the time average for 1957–94 by an overbar, Eq. (1) becomes

$$\lambda \overline{\Delta T} = \overline{Q} - \overline{F}. \qquad (2)$$

The heat content increase yields $\overline{F} = 0.32 \pm 0.15$ W m^{-2} (expressed per unit area of the entire world, not just the ocean surface).

Here ΔT is defined with respect to the steady-state climate for zero forcing. Sufficient measurements exist to estimate global-average temperature changes back to 1860, but the climate of that period was not a steady state, not least because anthropogenic greenhouse gases began to increase in the latter part of the eighteenth century. In fact, there has probably never been a steady-state climate, because solar output fluctuations and volcanism produce continual variations in radiative forcing on a shorter timescale than that required for the climate system to reach equilibrium.

TABLE 1. Radiative forcing difference $\overline{Q}'$ (W m^{-2}) between the periods 1957–94 and 1861–1900.

	-2σ	Central	$+2\sigma$
Greenhouse gases	1.24	1.38	1.51
Sulfate aerosols	−1.61	−1.01	−0.41
Solar irradiance changes	0.10	0.30	0.50
Volcanic aerosols	−0.49	−0.31	−0.12

As we do not know the global-average temperature for the steady state, we instead consider differences between the recent and an earlier period, as widely separated as possible in order to maximize the climate change signal. We take the difference between Eq. (2) for 1957–94 and a corresponding equation for the period 1861–1900, denoting the differences between the means for the two periods as $\overline{\Delta T}'$, $\overline{Q}'$, and $\overline{F}'$. Hence,

$$\lambda \overline{\Delta T}' = \overline{Q}' - \overline{F}'. \qquad (3)$$

Folland et al. (2001) calculated annual surface temperature anomalies, with uncertainties, by combining land- and ocean-based observations using an optimal averaging technique. From their data, the difference in global-average temperature between the two periods is $\overline{\Delta T}' = 0.335° \pm 0.033°$C, where the uncertainty (two standard deviations) was obtained assuming the two periods to be independent, and making allowance for serial correlation of annual values within each period.

No observations exist of past changes in radiative forcing, so this quantity must be estimated. We take into account the effects of greenhouse gases (carbon dioxide, methane, nitrous oxide, halocarbons, and tropospheric ozone), anthropogenic sulfate aerosols, solar variation, and volcanic aerosols (Table 1). Greenhouse gas and sulfate aerosol forcing are dominant and of opposite signs. The former is calculated using historical concentrations of the gases and formulas for radiative forcing; there is estimated to be a range of uncertainty of $\pm 10\%$ on the results (Ramaswamy et al. 2001).

The effect of sulfate aerosol is much less precisely known. The patterns of temperature change are sensitive to aerosol forcing. We derive limits for the forcing (Table 1) by comparison of the spatiotemporal patterns of temperature change in observations and experiments with the Hadley Centre AOGCM (HadCM3; Stott et al. 2000). The method (see Allen et al. 2002) assumes that the patterns simulated by

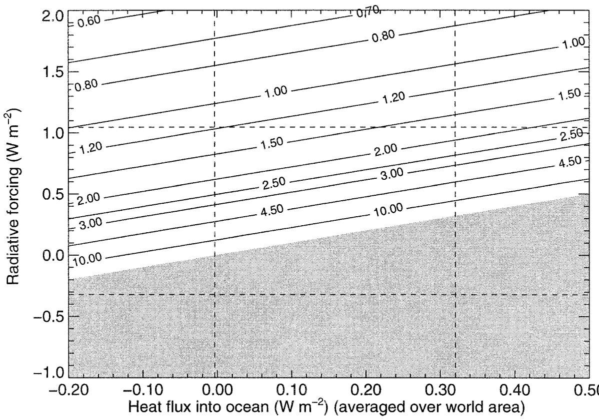

FIG. 1. Effective climate sensitivity $\Delta T_{2\times}$ as a function of radiative forcing $\overline{Q}'$ and heat flux $\overline{F}'$ into the ocean for the central value of surface temperature change $\overline{\Delta T}'$. The qualitative form of this function is the same for other choices of $\overline{\Delta T}'$ within its range of uncertainty, which is considerably less important than those in $\overline{Q}'$ and $\overline{F}'$. The dashed lines parallel to the axes indicate the $\pm 2\sigma$ intervals for $\overline{Q}'$ and $\overline{F}'$. The oblique lines of constant $\Delta T_{2\times}$ are straight because they apply to constant values of $\overline{Q}' - \overline{F}'$. In the shaded region, $\Delta T_{2\times}$ is negative; it is infinite on the boundary between the shaded and unshaded regions, which passes through the origin.

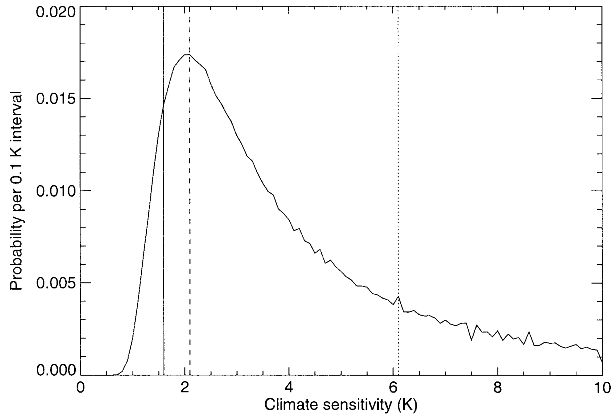

FIG. 2. Probability distribution for the effective climate sensitivity $\Delta T_{2\times}$, computed assuming $\overline{\Delta T}'$, $\overline{F}'$, and $\overline{Q}'$ to be normally distributed. The bin width is 0.1 K. The vertical solid line marks the lower bound of the 90% confidence interval (5th percentile), the vertical dashed line the modal value of $\Delta T_{2\times}$, and the vertical dotted line the median. Although the distribution is shown here only up to 10 K, the probability of larger values was accounted for in deriving the statistics and confidence interval.

the AOGCM are realistic, but does not depend on the model's forcing or climate sensitivity.

Solar output is thought to have increased in the early twentieth century, giving a positive contribution to $\overline{Q}'$, for which we take the range of Ramaswamy et al. (2001). Relative to the long-term average, there was a large amount of volcanism during recent decades, including the eruptions of Agung, El Chichon, and Pinatubo. Despite the eruption of Krakatoa, the late nineteenth century was on average less active, resulting in a negative contribution of volcanic forcing to $\overline{Q}'$ of -0.2 W m^{-2} (Andronova et al. 1999), or -0.4 W m^{-2} (Crowley 2000).

We obtain a $\pm 2\sigma$ interval for forcing change $\overline{Q}'$ of between -0.3 and $+1.0$ W m^{-2}, by combining the ranges of the various terms (Table 1), assuming the individual ranges to be normal $\pm 2\sigma$ intervals. We make this assumption on pragmatic grounds, as we lack knowledge of the probability density function of any of the terms, although we note that a tendency toward normality as one adds more terms is consistent with the central limit theorem.

To complete the calculation, information is needed about the average heat flux into the ocean during 1861–1900, in order to calculate $\overline{F}'$. In the absence of observational data, experiments both with AOGCMs and with simpler climate models (e.g., Stott et al. 2000; Forest et al. 2002) commonly assume that the climate was in a steady state at the starting point of their integrations, typically in the late nineteenth century. We investigate this assumption using the simple climate model of Raper et al. [1996, which implements Eq. (1) and calculates F with a one-dimensional upwelling-diffusion ocean model], as tuned by Cubasch et al. (2001) to reproduce results from a range of AOGCMs. In simulations with anthropogenic forcing alone, starting at the onset of substantial industrial emissions of greenhouse gases in the late eighteenth century, $\overline{F}$(1861–1900) lies in the range 0.06–0.10 W m^{-2}. In simulations with natural forcings alone (volcanoes and solar variation; Crowley 2000), starting in the year 1000, $\overline{F}$(1861–1900) is coincidentally also 0.06–0.10 W m^{-2}. It is positive because the climate is recovering from substantial negative forcing in preceding decades. Making the usual assumption that forcings can be combined linearly, these results suggest an estimate for $\overline{F}$(1861–1900) of 0.12–0.20 W m^{-2}, whereas a steady state would have $\overline{F}$(1861–1900) $= 0$. Treating this range as a normal $\pm 2\sigma$ interval and calculating the difference from $\overline{F}$(1957–94) obtained above, we obtain a $\pm 2\sigma$ interval for $\overline{F}'$ of 0.00–0.32 W m^{-2}.

3. Results

We calculate λ from Eq. (3) as a function of $\overline{\Delta T}'$, $\overline{F}'$, and $\overline{Q}'$, and convert it to $\Delta T_{2\times}$ using $Q_{2\times} = 3.71$ W m^{-2} (Myhre et al. 1998; Fig. 1). We compute the probability distribution of resulting values (Fig. 2), assuming $\overline{\Delta T}'$, $\overline{F}'$, and $\overline{Q}'$ to be independently and normally distributed with the standard deviations derived above, and ignoring the uncertainty of $\sim$1% in $Q_{2\times}$ (Myhre et al. 1998), which is negligible by comparison. The effect of internal (unforced) variability of the climate system on $\overline{F}'$ and $\overline{Q}'$ is also neglected, because estimates based on 1300 years of the HadCM3 control run show these fluctuations to be an order of magnitude smaller than the uncertainties. From the probability distribution of $\Delta T_{2\times}$ we obtain a 90% confidence interval, whose lower bound (the 5th percentile) is 1.6 K. The median is 6.1 K, above the canonical range of 1.5–4.5 K; the mode is 2.1 K.

A positive $\overline{F}$(1861–1900) implies that some of the twentieth-century warming is a committed response to previous forcing (Weaver et al. 2000). If the late nineteenth century is assumed to be a steady-state climate, such that $\overline{F}(1861\text{–}1900) = 0$, the 5th percentile of $\Delta T_{2\times}$ increases to 2.0 K. On the other hand, if the climate system were assumed always to be in steady state, that is, $\overline{F}' = 0$, the 5th percentile of $\Delta T_{2\times}$ would be 1.3 K. Use of a low-diffusivity ocean model might underestimate heat uptake, thus giving smaller $\Delta T_{2\times}$.

The 90% confidence interval for $\Delta T_{2\times}$ extends up to infinity, and beyond to negative values (cf. Fig. 1). Here $\Delta T_{2\times} < 0$ if $\overline{Q}' < \overline{F}'$, which means that heat flux into the ocean has increased by more than the radiative forcing. Negative $\Delta T_{2\times}$ is unphysical, because it implies that the unforced climate system would be unstable to any perturbation generated by internal variability. We infer that $\overline{Q}' < \overline{F}'$ should be regarded as implausible. With $\overline{Q}'$ only slightly greater than $\overline{F}'$, $\Delta T_{2\times}$ is extremely large. Such values can be excluded by paleoclimatic studies, which show that the climate sensitivity of the real world is of roughly the size indicated by GCMs (e.g., Hoffert and Covey 1992), but do not constrain it more tightly.

The dominant uncertainty in the calculation of climate sensitivity is clearly that pertaining to the estimates of radiative forcing, especially the aerosol forcing (cf. Forest et al. 2002; Knutti et al. 2002; Allen et al. 2002). While representing the state of current knowledge, the radiative forcing estimates we have employed are imprecise and undoubtedly incomplete in some respects. Some known negative radiative forcings have been omitted (stratospheric ozone depletion, aerosol from biomass burning, albedo change from land use change; Ramaswamy et al. 2001), whose inclusion would tend to raise the lower bound of $\Delta T_{2\times}$. Mineral dust and black carbon aerosol, also omitted, could give positive forcing (Ramaswamy et al. 2001). If we make an informal allowance for the possibility of substantial additional positive forcing by raising the upper bound of the sulfate aerosol forcing to zero, following Andronova and Schlesinger (2001), the 5th percentile of the climate sensitivity falls to 1.1 K. Although the HadCM3 simulations from which the sulfate aerosol forcing was derived did not include nonsulfate anthropogenic aerosols, these may have a somewhat similar geographical distribution to that of sulfate aerosols. To the extent that this is so, the sulfate aerosol forcing resulting from the method includes them as well; otherwise, their omission will be reflected in a greater forcing uncertainty.

We consider that the lower bound is an important constraint on climate sensitivity, because it is objectively derived, and independent of GCM results for $\Delta T_{2\times}$. Although the lower bound does not lead us to reject any of the AOGCMs used by Cubasch et al. (2001) in projections for the twenty-first century, it does exclude substantially smaller values. Improved understanding of physical processes of climate change and refinement of climate models is essential to reducing uncertainty in climate prediction. However, reducing the uncertainty on the inputs to the method described here offers an alternative route to obtaining better constraints on climate sensitivity. For example, with $\overline{Q}' = 0.8$ W m^{-2} and $\Delta T_{2\times} = 2.0$ K, if the ranges of uncertainty on $\overline{Q}'$ and $\overline{F}'$ were both $\pm 10\%$ (the same as the present uncertainty on greenhouse gas forcing), the 5%–95% confidence interval for $\Delta T_{2\times}$ from this method would be 1.7–2.3 K. A range as narrow as that would be a great improvement on the current state of knowledge.

Acknowledgments. We are grateful to Myles Allen, Tom Crowley, John Mitchell, Tom Wigley, Michael Schlesinger, John Antonov, Syd Levitus, and the anonymous reviewers for discussions, comments, and other help. Work at the Hadley Centre was supported by the U.K. Department of the Environment, Food and Rural Affairs under Contract PECD 7/12/37 and by the Government Meteorological Research contract. Sarah Raper was supported at AWI by HGF Strategiefonds Projekt 2000/13 SEAL.

REFERENCES

Allen, M. R., and Coauthors, 2002: Quantifying anthropogenic influence on recent near-surface temperature change. *Surv. Geophys.,* in press.

Andronova, N. G., and M. E. Schlesinger, 2001: Objective estimation of the probability density function for climate sensitivity. *J. Geophys. Res.,* **106,** 22 605–22 612.

——, E. G. Rozanov, F. Yang, M. E. Schlesinger, and G. L. Stenchikov, 1999: Radiative forcing by volcanic aerosols from 1850 to 1994. *J. Geophys. Res.,* **104,** 16 807–16 826.

Crowley, T. J., 2000: Causes of climate change over the past 1000 years. *Science,* **289,** 270–277.

Cubasch, U., and Coauthors, 2001: Projections of future climate change. *Climate Change 2001: The Scientific Basis,* J. T. Houghton et al., Eds., Cambridge University Press, 525–582.

Folland, C. K., and Coauthors, 2001: Global temperature change and its uncertainties since 1861. *Geophys. Res. Lett.,* **28,** 2621–2624.

Forest, C. E., P. H. Stone, A. P. Sokolov, M. R. Allen, and M. D. Webster, 2002: Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science,* **295,** 113–117.

Hoffert, M. I., and C. Covey, 1992: Deriving global climate sensitivity from paleoclimate reconstructions. *Nature,* **360,** 573–576.

Knutti, R., T. F. Stocker, F. Joos, and G. K. Plattner, 2002: Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature,* **416,** 719–723.

Levitus, S., J. I. Antonov, T. P. Boyer, and C. Stephens, 2000: Warming of the world ocean. *Science,* **287,** 2225–2229.

——, ——, J. Wang, T. L. Delworth, K. W. Dixon, and A. J. Broccoli, 2001: Anthropogenic warming of the Earth's climate system. *Science,* **292,** 267–270.

Myhre, G., E. J. Highwood, K. P. Shine, and F. Stordal, 1998: New estimates of radiative forcing due to well mixed greenhouse gases. *Geophys. Res. Lett.,* **25,** 2715–2718.

Ramaswamy, V., and Coauthors, 2001: Radiative forcing of climate change. *Climate Change 2001: The Scientific Basis,* J. T. Houghton et al., Eds., Cambridge University Press, 349–416.

Raper, S. C. B., T. M. L. Wigley, and R. A. Warrick, 1996: Global sea-level rise: Past and future. *Sea-Level Rise and Coastal Subsidence,* J. D. Milliman and B. U. Haq, Eds., Kluwer Academic, 11–45.

Senior, C. A., and J. F. B. Mitchell, 2000: The time dependence of climate sensitivity. *Geophys. Res. Lett.,* **27,** 2685–2688.

Stott, P. A., S. F. B. Tett, G. S. Jones, M. R. Allen, J. F. B. Mitchell, and G. J. Jenkins, 2000: External control of 20th century temperature by natural and anthropogenic forcings. *Science,* **290,** 2133–2137.

Weaver, A. J., P. B. Duffy, M. Eby, and E. C. Wiebe, 2000: Evaluation of ocean and climate models using present-day observations and forcing. *Atmos.–Ocean,* **38,** 271–301.

Wigley, T. M. L., P. D. Jones, and S. C. B. Raper, 1997: The observed global warming record: What does it tell us? *Proc. Natl. Acad. Sci.,* **94,** 8314–8320.