A new method for diagnosing radiative forcing and climate sensitivity

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[1] We describe a new method for evaluating the radiative forcing, the climate feedback parameter (W m^{-2} K^{-1}) and hence the effective climate sensitivity from any GCM experiment in which the climate is responding to a constant forcing. The method is simply to regress the top of atmosphere radiative flux against the global average surface air temperature change. This method does not require special integrations or off-line estimates, such as for stratospheric adjustment, to obtain the forcing, and eliminates the need for double radiation calculations and tropopause radiative fluxes. We show that for CO_2 and solar forcing in a slab model and an AOGCM the method gives results consistent with those obtained by conventional methods. For a single integration it is less precise but since it does not require a steady state to be reached its precision could be improved by running an ensemble of short integrations. *INDEX TERMS:* 1610 Global Change: Atmosphere (0315, 0325); 1620 Global Change: Climate dynamics (3309); 3359 Meteorology and Atmospheric Dynamics: Radiative processes. **Citation:** Gregory, J. M., W. J. Ingram, M. A. Palmer, G. S. Jones, P. A. Stott, R. B. Thorpe, J. A. Lowe, T. C. Johns, and K. D. Williams (2004), A new method for diagnosing radiative forcing and climate sensitivity, *Geophys. Res. Lett.*, *31*, L03205, doi:10.1029/2003GL018747.

1. Introduction

[2] The concepts of radiative forcing and climate sensitivity are commonly used in analysis of climate change simulated by general circulation models (GCMs), because they are useful in comparing the size of response by different models and to different forcings (greenhouse gases, aerosols, solar variability, etc.). A radiative forcing applied to the climate system must result in a climate change which tends to counteract the forcing; otherwise the system would be unstable. The climate sensitivity measures the size of the global average surface air temperature response.

[3] To be quantitative, let the imposed forcing be F (positive downwards), and the radiative response caused by the climate change be H (positive upwards), both being global averages (W m^{-2}), both initially zero. These heat fluxes are usually evaluated at the tropopause, because heat is rapidly exchanged within the troposphere and at the surface; the troposphere, surface and upper ocean thus constitute a tightly coupled "climate system". The stratosphere tends to equilibrate separately and on a timescale of only a few months, more quickly than this system.

[4] The net downward heat flux $N = F - H$ is the rate of increase of heat stored in the climate system. Since the heat capacity of the system resides overwhelmingly in the ocean, on interannual timescales N at the tropopause is practically equal to the rate of ocean heat uptake. By definition, a steady-state climate requires that $N = 0$, so that heat storage is not changing.

[5] It has been found from model experiments that in any given GCM the radiative response H is proportional to the global average surface air temperature change ΔT. We write $H = \alpha\Delta T$, where α is the climate response parameter, indicating the strength of the climate system's net feedback. It is assumed that the real climate system has some constant α (presently unknown). Different GCMs may have different α, but in any given GCM α is found to be roughly independent of both climate state and forcing. Recent analyses have provided more information about the limitations of this approximation [*Hansen et al.*, 1997; *Senior and Mitchell*, 2000; *Joshi et al.*, 2003]. If α is not constant, it is less useful for making projections.

[6] In a perturbed steady state $F = H = \alpha\Delta T \Rightarrow \Delta T = F/\alpha$. The "equilibrium climate sensitivity" $\Delta T_{2\times}^{\mathrm{eqm}}$ is the conventional measure of the climate system's response to forcing, defined as the steady-state ΔT due to a doubling of the CO_2 concentration. If this gives forcing $F_{2\times}$, $\Delta T_{2\times}^{\mathrm{eqm}}$ and α are simply related according to $\Delta T_{2\times}^{\mathrm{eqm}} = F_{2\times}/\alpha$—the smaller α, the larger $\Delta T_{2\times}^{\mathrm{eqm}}$.

[7] Two methods are commonly used to evaluate α. In method A, a climate model is run to a steady state with known forcing, and α is given by $F/\Delta T$. This method is practicable with a "slab" model (an atmosphere GCM coupled to a mixed-layer ocean), because such models take only 10–20 years to reach a steady state. Coupled atmosphere-ocean GCMs (AOGCMs), however, take millennia, making this method computationally very expensive.

[8] In method B, α is evaluated from any AOGCM state, not necessarily a steady state, using N as well as F and ΔT, according to $\alpha = H/\Delta T = (F - N)/\Delta T$. Evaluated this way, α is often expressed as the "effective climate sensitivity" $\Delta T_{2\times}^{\mathrm{eff}} \equiv F_{2\times}/\alpha$, to permit comparison with $\Delta T_{2\times}^{\mathrm{eqm}}$. It turns out that α is not always constant as the climate changes. For instance, in an experiment with constant $2 \times CO_2$ using the HadCM2 AOGCM, *Senior and Mitchell* [2000] found that $\Delta T_{2\times}^{\mathrm{eff}}$ increased from 2.7 K to 3.8 K over 830 years.

[9] In both methods, we need F to compute α, but F is not straightforward to obtain. The increase in net downward tropopause radiation caused by instantaneously imposing the forcing agent (e.g. doubling CO_2) is diagnosed by

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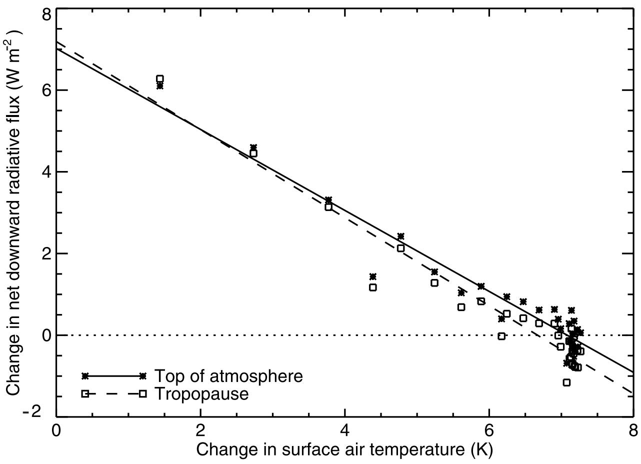

Figure 1. The evolution of global average net downward radiative flux with global average surface air (1.5 m height) temperature change in a HadSM3 experiment with constant $4 \times CO_2$. The dotted line is $N = 0$.

calculating the radiative fluxes in a model on each timestep with and without the forcing agent, using a simulation of a few years for a good estimate of the annual average. However, this is inadequate because not all the forcing to which the climate system responds appears instantaneously. For example, raising CO_2 concentration increases the rate that the stratosphere radiates heat to space. It adjusts by a temperature decrease taking a few months. This reduces the net downward tropopause radiation and hence the effective forcing due to the CO_2, but cannot be as easily diagnosed as the instantaneous forcing, and thus complicates the estimate of F. Some radiatively important processes have no instantaneous forcing, such as indirect and semi-direct effects [*Hansen et al.*, 1997] of aerosols, arising from changes in clouds.

2. Method

[10] We propose a new and simple method C for estimating F and α. In a climate experiment when the forcing agent has no interannual variation, we assume the forcing is constant on timescales of years or longer. Since

$$N = F - H = F - \alpha \Delta T, \tag{1}$$

if we plot the variation of $N(t)$ against $\Delta T(t)$ as the run proceeds in time t (using annual or longer-period means), we should get a straight line whose N-intercept is F and whose slope is $-\alpha$. A linear regression will give us both quantities without the need for any extra diagnostic techniques such as double radiation calculations. The ΔT-intercept will be F/α, equal to the equilibrium ΔT. If N is the net tropopause heat flux, F must be the tropopause forcing to which the climate system is responding on annual and longer timescales, because $N = F$ in the limit that $\Delta T \rightarrow 0$. In particular, F should include stratospheric adjustment, the indirect aerosol effect and others which do not cause an instantaneous radiative change. This leads us to suggest a practical distinction between a forcing and a feedback: Radiative forcing is a change in N brought about by the presence of the forcing agent, developing much more rapidly than the climate can respond (hence affecting the intercept). A climate feedback is a change in N which arises from the climate response to the forcing (hence affecting the slope) [cf. *Shine et al.*, 2003].

[11] *Shine et al.* [2003] have recently proposed an alternative method for evaluating forcing which is, in effect, to hold the climate system at the limit $\Delta T = 0$ in the presence of the forcing agent. In this situation $N = F$; so we expect that our method and theirs will give similar results for forcing.

[12] Our method can be applied to any experiment that has time-variation, using a slab model or an AOGCM. If α is constant, it is unnecessary to run the experiment to a steady state. For this method, unlike the usual method of diagnosing $\Delta T_{2\times}^{\mathrm{eqm}}$ from a slab experiment, it is the time-development before the steady state is reached which is of interest, not the steady state itself, because it is the time-variation which produces the straight line. If α is not constant, the points will not lie on a straight line. The variation of the slope provides a means of diagnosing the dependence of the feedbacks on climate state.

[13] We now compare the results of the different methods of estimating climate sensitivity and forcing using as examples experiments with the HadCM3 AOGCM [*Gordon et al.*, 2000] and the HadSM3 slab model [*Williams et al.*, 2001], which comprises HadAM3 (the atmosphere component of HadCM3) coupled to a "slab" ocean 50 m deep. Climate change in each model is calculated by subtracting the results of its own control experiment, which has constant atmospheric composition and a steady-state climate.

3. CO_2 Forcing in a Slab Model

[14] Starting from its control, an instantaneous quadrupling of CO_2 was imposed on HadSM3. It evolves towards a steady state over about 20 years. For reasons described later, we take the net downward radiative flux at the top of the atmosphere (TOA), instead of at the tropopause, as the net heat flux N into the climate system. We plot N against ΔT (Figure 1). The evolution starts at the top left, where N is large (initially equal to the forcing) and ΔT is small, and moves down and right as ΔT rises and N declines. There is scatter about a straight line resulting from the internally generated variability of the climate system. The steady state is reached when $N = 0$, at the ΔT-intercept. There is a cloud of points around this state, again because of internal variability. We exclude these points from the regression, because their relationship between N and ΔT may be different from that applying to climate change on decadal timescales.

[15] Regressing N against ΔT (Figure 1) for years 1–20 of the $4 \times CO_2$ experiment, we find that $F = 7.0 \pm 0.3$ W m^{-2}, implying $F_{2\times} = 3.5 \pm 0.2$ W m^{-2}. The stated uncertainty is the standard error from the regression (see below for discussion). The "standard" value of $F_{2\times} = 3.74 \pm 0.04$ W m^{-2} was determined using double radiation calculations in HadAM3 and an estimate of stratospheric adjustment. The regression is an easier method, and the results are statistically consistent. The agreement confirms that F from the regression method does include stratospheric adjustment (~ -1.0 W m^{-2} for $4 \times CO_2$), as postulated.

Table 1. Comparison of Results for the Climate Sensitivity from Various Experiments

M	Experiment	F W m^{-2}	α W m^{-2} K^{-1}	$\Delta T_{2\times}$ K
A	HadSM3 4 × CO_2	—	1.04 ± 0.01	3.59
C	HadSM3 4 × CO_2	7.0 ± 0.3	0.99 ± 0.07	3.8
C	HadCM3 2S yrs 1–90	3.9 ± 0.2	1.26 ± 0.09	3.0
C	HadCM3 4S yrs 1–90	7.5 ± 0.3	1.19 ± 0.07	3.1
B	HadCM3 4R yrs 1100–1200	—	0.91 ± 0.02	4.1
C	HadSM3 solar reduction	−1.2 ± 0.2	1.6 ± 0.4	2.4
A	HadSM3 solar increase	—	1.47 ± 0.05	2.5
C	HadSM3 solar increase	3.7 ± 0.3	1.3 ± 0.2	2.9
C	HadCM3 solar increase	4.2 ± 0.4	2.0 ± 0.3	1.9

The "M" column gives the method. Only method C gives a forcing; A and B require it as input. The standard $F_{2\times}$ = 3.74 ± 0.04 W m^{-2} was used to convert between $\Delta T_{2\times}$ and α.

[16] It makes little difference to the result for forcing if we evaluate N at the tropopause instead of the TOA. The intercept of 7.2 ± 0.3 W m^{-2} is statistically indistinguishable. This is because only the first few months of integration are affected by the stratospheric adjustment; once the stratosphere is in a steady state, the net heat fluxes at the tropopause and TOA must be equal.

[17] However, there is a difference in the final steady state: The net radiative flux at the tropopause in years 21–30 is −0.5 ± 0.2 W m^{-2}, while at the TOA it is indistinguishable from zero, being −0.1 ± 0.2 W m^{-2} (the uncertainty is the interannual standard deviation). The problem is that across the tropopause, unlike the TOA, there may be sensible and latent heat exchange in addition to radiative heat exchange. Apparently there is an increase in upward non-radiative heat flux of ~0.5 W m^{-2} at the tropopause arising from climate change, 7% of the forcing. For a correct estimate of α based on net radiative flux, we must evaluate N at the TOA. Our method thus does not use tropopause radiative fluxes, avoiding the need to diagnose the tropopause, an arbitrary and possibly systematically biased procedure [cf. *Shine et al.*, 2003].

[18] The regression slope for N against ΔT gives α = 0.99 ± 0.07 W m^{-2} K^{-1}. The uncertainty from the regression uses the RMS deviation in N (the dependent variable) from the fitted line to obtain an estimate of the uncertainty of the points. There are two possible problems with this. First, there is interannual correlation of variability so the points are not independent. This leads to an underestimate of uncertainty but not to a bias. Second, the choice of dependent variable is arbitrary. N and ΔT both have random noise, but regression assumes there is no uncertainty in ΔT. This tends to flatten the slope and underestimate its uncertainty. To gauge the size of the effect, we regress ΔT against N, obtaining a slope of −0.94 ± 0.06 K W^{-1} m^2, whose reciprocal is 1.06 ± 0.07 W m^{-2} K^{-1}. The effect is not serious. However, the product of the two slopes equals the square of the correlation coefficient, so the difference is greater when the points have more scatter. Correction for this could be made by a more elaborate procedure based on statistical properties of variability in the control experiment. Ordinary regression is adequate when statistical uncertainty is low.

[19] For the average of years 21–30, ΔT = 7.18 ± 0.05 K (interannual standard deviation). This gives an α more precise than the value from regression, but the two are statistically consistent (Table 1). By using the time-dependent part of the experiment, our method has obtained the same value for climate sensitivity as the normal method does from the final steady state. Its precision could be improved by using an ensemble of short integrations.

4. CO_2 Forcing in an AOGCM

[20] Experiments with an instantaneous doubling (experiment 2S, "S" for "sudden") and quadrupling (4S) of CO_2 have been run with HadCM3 starting from its control (Figure 2). These show similar behaviour to HadSM3 with 4 × CO_2 (Figure 1) but due to the larger heat capacity of the ocean they approach equilibrium more slowly. In the ninth decade of 4 × CO_2 HadCM3 has ΔT = 4.9 K, about 70% of the steady-state ΔT of HadSM3. The values of α from experiments 2S and 4S are consistent, but significantly larger than the α from HadSM3 with 4 × CO_2 forcing i.e. the climate sensitivity to CO_2 is smaller in HadCM3 than in HadSM3 (Table 1). The values of F for HadCM3 and HadSM3 are consistent.

[21] A longer HadCM3 4 × CO_2 experiment (4R, "R" for "ramp") was done whose 4 × CO_2 initial state was obtained by ramping up CO_2 from its control value. Following stabilisation of CO_2, ΔT rises for many centuries as the deep ocean slowly takes up heat [cf. *Senior and Mitchell*, 2000; *Voss and Mikolajewicz*, 2001], passing the steady-state for the 4 × CO_2 HadSM3 experiment. Averaged over years 1100–1200 the rate of temperature rise is ~10^{-3} K yr^{-1}. A clearer indication of continuing disequilibrium is that N = 0.7 W m^{-2} (Figure 2).

[22] The effective climate sensitivity calculated by method B also rises with time, similar to findings by [*Senior and Mitchell*, 2000] for HadCM2. Averaged over years 1100–1200, $\Delta T_{2\times}^{\mathrm{eff}}$ = 4.1 ± 0.1 K (uncertainty is the interdecadal standard deviation).

[23] When we plot N against ΔT for experiment 4R (Figure 2) we find they are not linearly related. By analogy

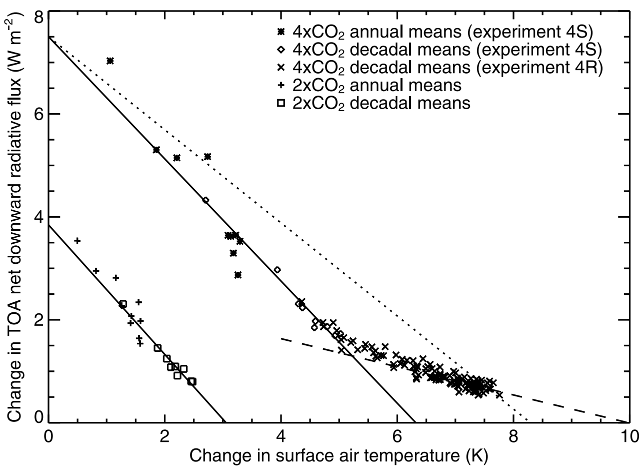

Figure 2. The evolution of global average TOA net downward radiative flux with global average surface air (1.5 m height) temperature change in HadCM3 experiments with fixed 2 × CO_2 (2S) and 4 × CO_2 (4S and 4R). The dashed line is a regression for years 500 onwards of experiment 4R, and the dotted line illustrates the calculation of $\Delta T_{2\times}^{\mathrm{eff}}$ for year 1000.

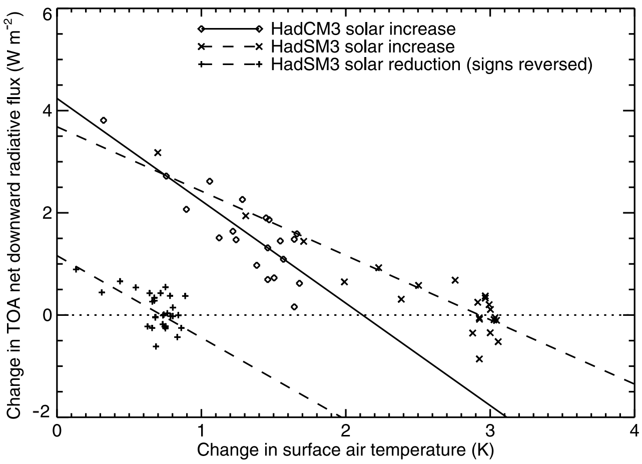

Figure 3. The evolution of global average TOA net downward radiative flux with global average surface air (1.5 m height) temperature change in HadSM3 and HadCM3 experiments with modified solar irradiance. The dotted line is $N = 0$.

to $\alpha = H/\Delta T$, one can define a "differential climate response parameter" $\alpha_{\mathrm{diff}} \equiv \mathrm{d}H/\mathrm{d}T = -\mathrm{d}N/\mathrm{d}T$ for constant F. Hence α_{diff} is (minus) the slope of the tangent to the N versus ΔT curve, and measures the feedbacks for small climate changes with respect to the current state. The usual climate response parameter α is (minus) the slope of a straight line between the starting point (0, F) and the present point (ΔT, N). It indicates the average strength of the feedbacks that occurred during the climate change.

[24] In Figure 2 we have fitted a straight line for years 500 onwards of experiment 4R; its slope gives $\alpha_{\mathrm{diff}} = 0.27 \pm 0.04$ W m^{-2} K^{-1}, considerably smaller than α (cf. Table 1). The physical mechanisms responsible for this difference are under investigation. The clear-sky longwave component of N varies linearly with ΔT; deviations from linearity are found in the clear-sky shortwave and in the cloud feedbacks.

[25] Since α_{diff} is not constant, we cannot reliably predict $\Delta T_{2\times}^{\mathrm{eqm}}$ without running the experiment out to a steady state. However, we can estimate it by extrapolating to $N = 0$ (dashed line in Figure 2). This gives $\Delta T = 10.0$ K, indicating $\Delta T_{2\times}^{\mathrm{eqm}} \simeq 5$ K, which should be taken as a lower limit, since the slope may show a continuing tendency to flatten. It is evident from the slow rate of temperature rise in the later part of the experiment that it would take a long time to reach this steady state. As a result, an exponential fit to the ΔT timeseries to obtain $\Delta T_{2\times}^{\mathrm{eqm}}$ [*Voss and Mikolajewicz*, 2001] might be relatively poorly constrained.

5. Solar Forcing

[26] Variations in solar irradiance could produce a significant radiative forcing, with magnitudes on decadal and century timescales estimated to be a few tenths W m^{-2}. We have undertaken three experiments (Figure 3) to evaluate the climate sensitivity to solar forcing: (1) Reduction of the solar irradiance in HadSM3 by 0.55%, at the upper end of the range of estimates for the difference in irradiance between the Maunder minimum and the present day. (2) Increase to the solar irradiance in HadSM3 by ten times its anomaly for 1989 from the timeseries of *Lean et al.* [1995], in which 1989 has the largest value. (3) As (2), imposed on HadCM3.

[27] Regression of N against ΔT in the HadSM3 experiments gives values $\sim$1.5 W m^{-2} K^{-1} for α (Table 1), in agreement with the value calculated from the steady-state warming. These values are significantly larger than for CO_2, but such a size of difference is consistent with other studies [*Hansen et al.*, 1997; *Joshi et al.*, 2003]. However, the HadCM3 $\alpha = 2.0 \pm 0.3$ W m^{-2} K^{-1} is larger still, suggesting a climate sensitivity about 60% smaller than the HadCM3 sensitivity to CO_2. It is also clear from Figure 3 that the extrapolation of the HadCM3 experiment will give a smaller warming than that realised in the corresponding HadSM3 experiment.

6. Conclusions

[28] We have shown that for CO_2 and solar forcing in HadSM3 and HadCM3 our new method gives results consistent with those obtained by conventional methods for forcing and climate sensitivity. In HadCM3 we find markedly different sensitivity to these two forcings, and intend to investigate other kinds. Forcings which are geographically localised, notably those due to aerosols, are of particular interest because their distribution may affect the relative importance of various climate feedback mechanisms and thus the climate sensitivity [*Shine et al.*, 2003].

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