An alternative to radiative forcing for estimating the relative importance of climate change mechanisms

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Abstract. Radiative forcing is widely used to measure the relative efficacy of climate change mechanisms. Earlier general circulation model (GCM) experiments showed that the global-mean radiative forcing could be used to predict, with useful accuracy, the consequent global-mean surface temperature change regardless of whether the forcing was due to, for example, changes in greenhouse gases or solar output. More recent experiments indicate that for changes in absorbing aerosols and ozone, the predictive ability of radiative forcing is much worse. Building on a suggestion from Hansen and co-workers, we propose an alternative, the "adjusted troposphere and stratosphere forcing". We present GCM calculations showing that it is a significantly more reliable predictor of this GCM's surface temperature change than radiative forcing. It is a candidate to supplement radiative forcing as a metric for comparing different mechanisms and provides a framework for understanding the circumstances in which radiative forcing is less reliable. *INDEX TERMS:* 1610 Global Change: Atmosphere (0315, 0325); 1620 Global Change: Climate Dynamics (3309). **Citation:** Shine, K.P. et al., An alternative to radiative forcing for estimating the relative importance of climate change mechanisms. *Geophys.Res.Lett.* (submitted)

1. Introduction

The Intergovernmental Panel on Climate Change (IPCC) [*IPCC,* 2001] and others have extensively used radiative forcing of climate change as the most basic way of comparing the importance of different causes of climate change. Radiative forcing is defined as the change in the irradiance at the tropopause following, for example, an increase in carbon dioxide concentration or a change in solar output. If all other parameters are held fixed it is termed the instantaneous radiative forcing. However, radiative forcing has a generally greater utility if the relatively fast process of stratospheric temperature adjustment, resulting directly from the imposition of the forcing, is taken into account [e.g. *Hansen et al.,* 1997*; IPCC,* 1995]. This is normally referred to, in shorthand, as the "adjusted radiative forcing". The global-mean forcing will be denoted ΔF_a.

There is an approximate relationship between ΔF_a and the global-mean equilibrium surface temperature response, ΔT_s, such that

$$\Delta T_s \approx \lambda \Delta F_a,$$

where λ is a climate sensitivity parameter. One of the main motivations for comparing ΔF_a amongst climate change mechanisms, rather than ΔT_s, is that the absolute value of λ is poorly known. Different general circulation models (GCMs) yield a spread of values of λ ranging from about 0.4 to 1.2 K $(Wm^{-2})^{-1}$, much of the uncertainty being due to intermodel differences in cloud feedbacks [e.g. *IPCC*, 2001]. Hence, differences in predictions of ΔT_s amongst models could result from differences in ΔF and/or λ.

However, a wide range of GCM experiments have indicated that in any one model, λ is approximately constant whether ΔF_a is caused by, for example, changes in greenhouse gas concentrations, solar output or scattering aerosols [e.g. *Boer and Yu*, 2003; *Hansen et al.*, 1997; *IPCC*, 1995; *Joshi et al.*, 2003, *Rotstayn and Penner*, 2001]. Consequently, in assessing the relative size of mechanisms that initiate climate change, it is more instructive to compare ΔF_a than ΔT_s. There were other reasons that motivated the adoption of ΔF_a as a measure of climate change. Calculation of ΔT_s requires multidecadal simulations of a GCM that are computationally demanding compared to the calculation of ΔF_a. This lower computational demand means that calculations of ΔF_a can use more detailed radiative transfer schemes and more effectively search the parameter space of uncertainties in the specification of changes in atmospheric composition.

Recently, the general utility of radiative forcing has come to be questioned, because GCM experiments have shown that in a given model, for certain climate change mechanisms, λ departs significantly from its value for a change in carbon dioxide concentration. Examples of such mechanisms are changes in absorbing aerosols [*Cook and Highwood*, 2003; *Hansen et al.*, 1997] and changes in upper tropospheric or lower stratospheric ozone [*Joshi et al.*, 2003; *Stuber et al.*, 2001]. In the case of absorbing aerosols, in some circumstances ΔF_a fails to predict even the *sign* of the consequent surface temperature change.

2. Alternatives to the adjusted radiative forcing

Recently, *Hansen et al.* [2002] proposed an alternative to ΔF_a, which they termed the "fixed sea surface temperature forcing"; the global-mean is denoted ΔF_{sst}. ΔF_{sst} is calculated in a GCM by holding sea-surface temperature (SST) fixed; in addition to allowing for stratospheric temperature adjustment, it allows for changes in tropospheric lapse rate and other characteristics such as cloud cover and stratospheric water vapour, provided these processes are not driven by sea surface temperature change. Once the model has been run to equilibrium, ΔF_{sst} can be diagnosed at any level within the atmosphere, since it is a constant with height, although the individual energy terms contributing to ΔF_{sst} will vary with height. So at the top of the atmosphere it is a purely radiative change, while at the surface it can include changes in latent and sensible heat fluxes, in addition to the radiative changes.

From a limited number of tests, *Hansen et al.* [2002] concluded that ΔF_{sst} was no better at predicting surface temperature change in their model than ΔF_a. However, we were inspired to pursue this idea for two reasons. First, *Hansen et al.* did not test its performance for causes of modelled climate change for which ΔF_a has been shown to be least reliable. Secondly and, as we will show, crucially, it is not immediately obvious whether the land surface temperatures should be held fixed in addition to the SST. Here we explore a modified form of the *Hansen et al*. suggestion. Although "*fixed surface temperature forcing"* would follow their terminology, we prefer to stress what is being changed (in the spirit of the conventional adjusted forcing) rather than what is being kept fixed. Hence we call our new forcing, the (global-mean) *adjusted troposphere and stratosphere forcing,* ΔF_{ats}. We rationalise this choice as the temperature responses over land and ocean surfaces are clearly related, although our ultimate justification is the performance of ΔF_{ats} as a predictor of our GCM's ΔT_s, compared to ΔF_{sst}.

Although the computation of ΔF_{ats} and ΔF_{sst} requires a GCM, in contrast to the calculation of ΔF_a, there are a number of reasons which support an exploration of their use, some of which have already been pointed out by *Hansen et al.* [2002]. First, GCMs are now much more widely available and amenable to integration on what are relatively modest computer resources. Second, the use of fixed (sea) surface temperatures means that relatively short integrations are required, of a few years, compared to a few decades when calculating ΔT_s; in principle, this lower computational demand could allow more detailed radiative transfer calculations to be performed or the parameter space of uncertainty to be better explored. Third, the calculation of ΔF_a requires the specification of an appropriate tropopause position and the use of some method for computing the stratospheric temperature change (and these temperature changes are restricted to those driven by radiative processes). There is no obvious choice of tropopause position for the purpose of radiative forcing calculations in a GCM. It has been shown that in a 1-D radiative convective model, the stratospheric temperature change and the ability of ΔF_a to predict ΔT_s are significantly sensitive to this choice [*Forster et al.*, 1997]. For the calculation of latitudinally-resolved ΔF_a, the fixed-dynamical heating approximation has often been used as a relatively crude method of estimating stratospheric temperature change. Fourth, one important class of forcings, the aerosol indirect and semi-direct forcings, already generally use (and in the case of the second indirect and semi-direct effects, requires) GCMs for the calculation of global-scale forcings.

3. Model experiments

To illustrate the performance of ΔF_{ats} we use the Reading Intermediate GCM (IGCM). This is a global spectral model run at T21 (approximately 6^o) horizontal resolution with 22 vertical levels. It is "intermediate" in the sense that the physical parameterisations (radiation, clouds, surface flux) are typical of what would have been state of the art in the 1980s; their relatively lower computational demand, compared to current state of the art models, allows a wider set of calculations to be performed. When the ocean surface temperature change is enabled, it is calculated using a standard mixed layer ocean with specified oceanic heat fluxes to ensure that the model's geographical and seasonal variations in SST in the control run follow the observed SST. The IGCM's response to a range of climate perturbations has been shown to be encouragingly similar to the much more advanced GCM, ECHAM4 [*Joshi et al.*, 2003] when scaled to each model's response to CO_2 changes.

The calculations presented here are based mainly on model integrations from a study of the semi-direct aerosol forcing by *Cook and Highwood* [2003] which used a 2m mixed layer ocean to speed the approach to equilibrium. Two additional calculations examining the impact of ozone changes from *Joshi et al.* [2003], using a 25m mixed layer ocean, are also included. These use a slightly different version of the IGCM that possesses a different λ; the results presented here have been rescaled so the two sets of results have the same λ for increases in carbon dioxide concentration. ΔF_{ats} and ΔF_{sst} are calculated using a 5-year integration of the model with spatially varying sea and land surface temperatures taken from a monthly mean, annually-repeating observed climatology. ΔF_a is calculated using a fixed dynamical heating model using atmospheric characteristics taken from the IGCM control run. ΔT_s is calculated from the temperature change using the mixed-layer ocean after 30 years.

Results are presented for a globally uniform lower tropospheric aerosol layer with a mid-visible single scattering albedo, ω, ranging from 0.8 to 1.0 in steps of 0.05, plus a further case with a mid-tropospheric aerosol layer with ω=0.8. The mid-visible optical depth is 0.2. In addition, a standard $2xCO_2$ integration and one with a spectrally constant increase in total solar irradiance of 2% are included. In the ozone experiments, two cases are examined, for which λ departs most markedly from its value for the increased CO_2 in the IGCM. In the upper tropospheric ozone case, ozone is increased in the three layers below the climatological tropopause (approximately 90 - 220 hPa in the tropics and 210 - 440 hPa in high latitudes); in the lower stratospheric case, ozone is increased between 20 and 90 hPa. In both cases, the increase in ozone is chosen so as to give a ΔF_a of 1 Wm^{-2}.

ΔF_{sst} and ΔF_{ats} differ from ΔF_a in that these forcings cannot be calculated with arbitrary precision because of the internally generated variability in the global mean energy budget even with fixed surface temperatures. From 35 years of simulation, the standard deviation in ΔF_{ats} is 0.3 Wm^{-2}. This is a somewhat lower precision than that stated by *Hansen et al.* [2002] who indicate a 0.1 Wm^{-2} precision can be achieved with a GCM run of a few years. Note that the level of precision in the IGCM would preclude its use for calculating small forcings, such as those resulting from realistic changes in halocarbon concentrations.

The ΔF_{sst} and ΔF_{ats} results presented here are computed from the change in surface energy balance; the average difference with the value computed at the top of the atmosphere is 0.15 Wm^{-2}. We choose the surface value here, as the analysis of the change in the non-radiative components of the flux is instructive in understanding the changes.

4. Results

Table 1 shows the results for the 10 integrations reported here. The key results are in the final three columns. λ varies widely over almost two orders of magnitude, and in one case is negative, results similar to those of *Hansen et al.* [1997]. λ_{sst} (i.e. the climate sensitivity parameter calculated using ΔF_{sst} instead of ΔF_a) can be seen to be better behaved. (Not all results are presented for this parameter, as its performance is inferior to λ_{ats}.) For the more simple forcings in the top three rows of Table 1 (changes in carbon dioxide, solar output and absorbing aerosols), there is even closer agreement between λ_{sst} than λ. However, the variations remain large, with a factor of 6 difference between the smallest and largest value of λ_{sst}. The final column, λ_{ats}, shows remarkable agreement amongst all experiments presented here, with a spread of values about the mean value of 0.46 $K(Wm^{-2})^{-1}$ of only about 10%. A corollary of this agreement is that ΔF_{ats} predicts the IGCM's critical ω (i.e. the value at which the aerosol causes a zero change in ΔT_s) almost perfectly, in contrast to ΔF_{sst} and ΔF_a (Fig. 1) which are zero at values of ω for which ΔT_s is not zero; this leads to large and/or negative values of λ and λ_{sst} in the vicinity of this ω. Also of note is that λ_{ats} performs even better than λ_{sst} for the three simpler forcings.

In order to help understand the impact of absorbing aerosols we start by noting that, to first order, the global mean energy balance of the troposphere can be conceptualised as that required to maintain the observed temperature lapse rate. Net radiative cooling of the troposphere is balanced by the input of energy from the surface in the form of sensible and latent heat. Hence, the presence of absorbing aerosol in the troposphere reduces the net radiative cooling and lowers the "demand" for fluxes of energy from the surface. The decrease in latent heat flux with decreasing ω is shown in Fig. 2. Also shown in Fig. 2 is the change in downwelling longwave radiation. Since the aerosol itself affects only the shortwave radiation (in this model) any change in downwelling radiation must be driven by a change in the atmosphere as a result of the aerosol. The decrease in cloud amount as a result of the absorbing aerosol (the so-called semi-direct effect, which is at the heart of the lack of constancy of λ for absorbing aerosols) is clearly being captured in the calculation of ΔF_{ats} when it cannot, of course, impact on the calculation of ΔF_a.

The reason for the better behaviour of λ_{ats} over λ_{sst} is presumably because calculations of ΔF_{sst}, by allowing the land surface temperature to vary, will have some element of climate response included. However, this land surface temperature change may not represent in pattern or size, the eventual land surface response when SSTs are allowed to vary.

5. Conclusions

This paper has shown that ΔF_{ats} is an excellent predictor of the IGCM's surface temperature change and is greatly superior to the "standard" adjusted radiative forcing. Part of this improvement is due to the fact that it can allow changes in atmospheric parameters that are specific to a particular forcing; the so-called semi-direct forcing whereby absorbing aerosols reduce GCM cloudiness is an obvious example.

Another part of this improvement is because, unlike ΔF_a, ΔF_{ats} does not rely on an arbitrary definition of tropopause position. *Forster et al.* [1997] showed that in a 1-D radiative-convective model, the top of the convection was the ideal choice for tropopause position to calculate ΔF_a; while such a level can be readily identified in a 1-D model, this is not the case either in a GCM or in reality. In the calculation of ΔF_{ats}, there is no need to make such a choice as the model automatically distinguishes between altitudes where the temperature change is strongly constrained by surface temperature change, and altitudes under strong radiative control that can, therefore, undergo significant "stratospheric" adjustment.

The calculations presented here, showing the utility of ΔF_{ats}, cannot yet be regarded as conclusive without further work, most particularly comparison with results from other GCMs and for a wider range of radiative forcing mechanisms; a particular issue concerns spatially inhomogeneous forcings, where deviations of λ from the value for homogeneous forcings are driven by the geographical distributions of feedbacks [*Boer and Yu,* 2003]. Nevertheless the results are strongly suggestive of an improved metric. If further experiments established that ΔF_{ats} is a uniformly good predictor in different GCMs, other issues would remain. It would be necessary to establish whether it is the absolute value of ΔF_{ats} that should be compared between different models, or its value relative to, for example, a model's ΔF_{ats} for homogeneous changes in carbon dioxide.

The results also indicate the way forward to an improved simple conceptual model of climate change. A widely used model separates forcing from temperature response based on ΔF_a. This has led to a debate as to whether forcings such as the indirect and semi-direct aerosol effects are true forcings as they lead to changes in atmospheric parameters and require the use of a GCM to calculate [e.g. *IPCC*, 2001; *Rotstayn and Penner*, 2001]. The ΔF_{ats} framework is clearer as it distinguishes between forcings that change atmospheric parameters in the absence of surface temperature changes, and climate feedbacks that are ultimately mediated by the surface temperature change.

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Table 1. Global mean surface temperature change ΔT_s (in K), global mean forcing ΔF (in Wm^{-2}) and climate sensitivity parameter λ (in K $(Wm^{-2})^{-1}$) for a range of climate model experiments shown in the first column[a].

Expt	ΔT_s	ΔF_a	ΔF_{sst}	ΔF_{ats}	λ	λ_{sst}	λ_{ats}
$2xCO_2$	1.9	3.8	4.3	4.3	0.50	0.44	0.44
+2% S_o	1.9	4.9	4.9	4.2	0.39	0.39	0.45
L1.0	-1.7	-4.7	-4.6	-4.1	0.36	0.37	0.42
L0.95	-0.6	-3.0	-2.2	-1.3	0.20	0.28	0.48
L0.9	0.6	-1.4	0.3	1.4	-0.43	1.99	0.44
L0.85	1.8	0.14	2.6	4.0	12.9	0.7	0.46
L0.8	2.9	1.6	4.8	6.8	1.8	0.61	0.43
M0.8	1.2	5.0	-	2.7	0.24	-	0.45
UTO_3	0.3	1	-	0.68	0.30	-	0.48
LSO_3	0.66	1	-	1.26	0.66	-	0.52

[a]The subscript 'a' denotes (stratospheric temperature) adjusted radiative forcing; subscripts 'sst' and 'ats' refer to the GCM calculated forcing using fixed sea surface temperature and adjusted troposphere and stratosphere respectively. For the experiments, S_o is total solar irradiance, L refers to aerosols in the lower troposphere, M to aerosols in the middle troposphere; the value following L and M is the mid-visible single scattering albedo of that aerosol. UTO_3 refers to changes in upper tropospheric ozone, LSO_3 refers to changes in lower stratospheric ozone. Not all "sst" results are presented because of its inferior performance compared to "ats".

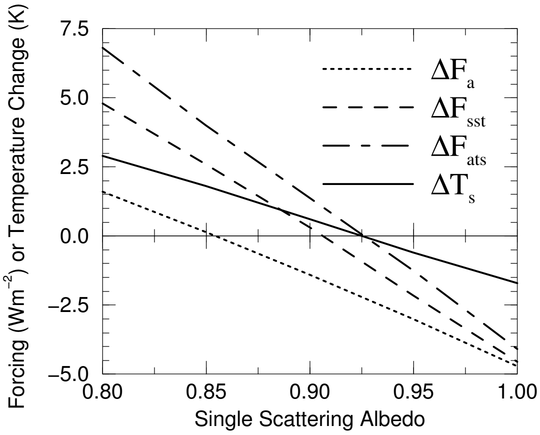

Figure 1: Adjusted radiative forcing, fixed sea-surface temperature forcing and adjusted troposphere and stratosphere forcing (in Wm^{-2}) as a function of mid-visible single scattering albedo. The GCM surface temperature change (in K) is also shown. All values are global means.

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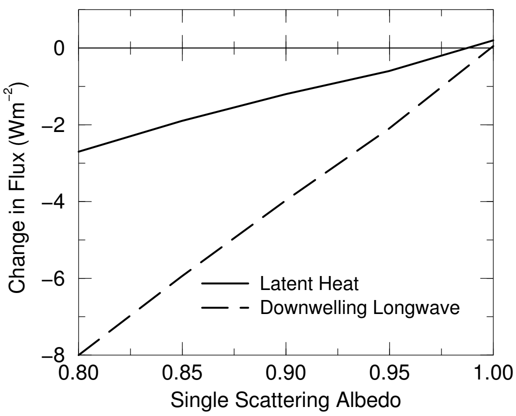

Figure 2: Change in latent heat flux and downwelling longwave irradiance (in Wm^{-2}) at the surface as a function of mid-visible single scattering albedo, for the adjusted troposphere and stratosphere forcing. All values are global means.

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