Tropospheric adjustment induces a cloud component in CO$_2$ forcing

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Abstract

The radiative forcing of CO$_2$ and the climate feedback parameter are evaluated in several climate models with slab oceans by regressing the global-mean annual-mean top-of-atmosphere radiative flux against the global-mean annual-mean surface air temperature change $\Delta T$ following a doubling of atmospheric CO$_2$ concentration. The method indicates that in many models there is a significant rapid tropospheric adjustment to CO$_2$ leading to changes in cloud, and reducing the effective radiative forcing, in a way analogous to the indirect and semi-direct effects of aerosol. By contrast, in most models the cloud feedback is small, defined as the part of the change which evolves with $\Delta T$. Comparison with forcing evaluated by fixing sea surface conditions gives qualitatively similar results for the cloud components of forcing, both globally and locally. Tropospheric adjustment to CO$_2$ may be responsible for some of the model spread in equilibrium climate sensitivity, and could affect time-dependent climate projections.

1. Introduction

Radiative forcing $F$ is the change in the net downward heat flux (W m$^{-2}$) into the climate system caused by an agent of climate change, such as an alteration to the atmospheric concentration of carbon dioxide or aerosol. The concept is useful because the magnitude of climate change that would be caused by different agents can often be estimated, without running a climate model, from their radiative forcings. Precise radiative forcings can be evaluated from detailed radiative calculations with less computation than required for a climate simulation, and can be compared without the complication of climate variability.

The magnitude of climate change is quantified by the change in global-mean surface air temperature change $\Delta T$ from the initial state. General circulation model (GCM) experiments show that $\Delta T \propto F$ in a steady state. GCMs give a wide range of values (Cubasch et al., 2001)
for the constant $\alpha$ (W m$^{-2}$ K$^{-1}$) in the formula $F = \alpha \Delta T$. This $\alpha$ is called the “climate feedback parameter” and is sometimes quoted equivalently as the “climate sensitivity parameter” $1/\alpha$ (K W$^{-1}$ m$^2$). Several analyses have been undertaken which use observed mean climate or climate change to constrain the climate sensitivity parameter of the real world (e.g. Gregory et al., 2002; Murphy et al., 2004; Forster and Gregory, 2006; Forest et al., 2006; Hegerl et al., 2006).

Thus, steady-state global-mean climate response is determined jointly by forcing $F$ and feedback $\alpha$. This factorisation is useful to the extent that feedback is model-dependent but the same in a given model for all forcing agents, while forcing is model-independent but depends on the forcing agent (Hansen et al., 2005).

The relationship between $\Delta T$ and $F$ is an expression of the energy balance of the climate system. As the climate changes in response to the forcing agent e.g. by warming up when the CO$_2$ concentration has been raised, it produces a radiative response $H = \alpha \Delta T$ which opposes the imposed $F$, so that the net heat flux into the climate system $N = F - H = F - \alpha \Delta T$. In the perturbed steady state $N = 0 \Rightarrow F = \alpha \Delta T$, since no more heat is being absorbed. ($N$ is most conveniently measured as the net downward radiative flux at the top of the atmosphere, TOA.)

The separation into forcing and feedback may appear to be arbitrary, since both the radiative forcing $F$ and the radiative response $H = \alpha \Delta T$ are consequences of imposing the forcing agent. The radiative forcing is sometimes defined as global-mean net downward change in radiative flux at the tropopause caused instantaneously when the forcing agent is imposed, but this definition is thought inadequate in the most important cases of

- CO$_2$, since raising the CO$_2$ concentration increases the rate at which the stratosphere radiates heat. It adjusts by a temperature decrease taking a few months. This reduces the net downward tropopause radiation.

- aerosol, since only the “direct” effect is instantaneous. Aerosols in the atmosphere also alter the occurrence and optical properties of clouds, which in turn have an “indirect” radiative effect, but these changes are not instantaneous.

Special treatments have been developed for these and other cases, in order to include in $F$ those effects which are considered to be part of the forcing. This is done partly because the resulting values for $F$ are more nearly model-independent and hence useful for comparative studies of the climatic effects of different forcings (e.g. Shine et al., 2003; Hansen et al., 2005).

The arbitrariness of what to count as forcing in a given model is a consequence of looking only at steady-state changes, from which it is not possible to distinguish between effects driven directly by the forcing agent and those associated with global climate change. Suppose that two different definitions of CO$_2$ forcing (for instance, including or excluding stratospheric adjustment) give values of $F$ and $F'$ for a doubling of CO$_2$. The GCM is run to steady state with doubled CO$_2$, and the diagnosed temperature change $\Delta T_{2x}$, called the “equilibrium climate sensitivity”, is used to calculate two different values of the feedback parameter $\alpha = F/\Delta T_{2x}$ and $\alpha' = F'/\Delta T_{2x}$. Using either of these values, we predict that twice the forcing will give twice the temperature change: $\Delta T = (2F)/\alpha = (2F')/\alpha' = 2\Delta T_{2x}$. Both forcing and feedback can be multiplied by any factor in this way without changing our predictions for the steady state.

The distinction of forcing and feedback is not arbitrary if we assume that the concepts should apply to time-dependent climate change, since different choices of $(F, \alpha)$ giving the same steady-state $\Delta T$ lead to different predictions of the heat flux into the climate system for a given $\Delta T$: $N = F - \alpha \Delta T = (F/F')(F' - \alpha' \Delta T) \neq F' - \alpha' \Delta T$. Since we regard “forcing”
as the change in $N$ caused directly by the forcing agent, and climate feedback as being the change in $N$ due to the response of the system to the forcing, time-dependence therefore suggests a practical way to evaluate forcing: $F$ is the limit of $N$ as $\Delta T \rightarrow 0$, where $\Delta T$ is measuring “climate response”. This defines the forcing as the net heat flux into the climate system caused by the forcing agent without any climate response yet having occurred. The distinction between forcing and feedback is thus made according to timescale. The radiative response of the climate system is distinguished from the forcing because the former develops over longer timescales, of years if only the mixed layer of the ocean is considered, or centuries if the full ocean.

Gregory et al. (2004) showed that when $N$ is plotted against $\Delta T$ for a fixed forcing of the HadSM3 slab climate model (Hadley Centre slab climate model version 3, see also Section 2a), a straight line gives a good fit, as would be expected from $N = F - \alpha \Delta T$, if $\alpha$ is constant. (In a more complex system, $\alpha$ might depend on climate and hence not be constant.) The $N$-intercept is the forcing $F$, following the prescription $F = \lim_{\Delta T \rightarrow 0} N$. Forcings such as stratospheric adjustment to CO$_2$ and the indirect forcing by aerosols are effects which develop quickly (over months or less) after the forcing agent is added. Consequently if means over long enough periods are used for the regression, $F$ includes these influences on forcing. Although stratospheric adjustment and aerosol forcing develop over periods of less than a year, it is convenient to use periods of a year or longer for the analysis so that the seasonal cycle does not have to be considered. Gregory et al. used annual or decadal means.

In this paper we pursue these ideas somewhat further. We argue that this way of distinguishing forcing and feedback is useful in explaining differences in climate response among GCMs, and consistent with the recently proposed methods of Hansen et al. (2002, 2005) and Shine et al. (2003) for evaluating the forcing.

2. Forcing and feedback diagnosed from time-dependent climate change

The climate radiative response arises from a number of feedback processes. A conventional decomposition $H = \sum_i H_i$ at the TOA is into longwave and shortwave radiation (terrestrial and solar), and clear-sky and cloud (Cess and Potter, 1988). Gregory et al. (2004) did not show components of $H$, but if $H \propto \Delta T$, it is reasonable to expect that the individual $H_i \propto \Delta T$ as well, so that $N_i = F_i + Y_i \Delta T$, where $F_i$ are the components of forcing and $Y_i$ the individual climate feedback parameters. (Note that we choose to write $Y_i$ with the opposite sign to $\alpha$, so that $\alpha = -\sum_i Y_i$, because that suits their physical interpretation below. The notation is summarised in Table 1.) A linear dependence of the $N_i$ on $\Delta T$ has been shown by Stowasser et al. (2006) and Lambert and Faull (2007) for GCMs forced with increased insolation, and by Forster and Gregory (2006) for observed changes in TOA radiation in recent years.

In GCMs the clear-sky components are obtained from a diagnostic radiative calculation (i.e. not affecting model evolution) with clouds assumed to be absent, and the cloud components as the difference between the radiation calculations with and without clouds. The sum of clear-sky and cloud components is called “all-sky”. The cloud components of $N$ are often referred to as “cloud radiative forcing”, a term we prefer to avoid because it is confusing. Because clouds are cooler than the surface, especially high clouds, they have a greenhouse effect, meaning that TOA outgoing longwave radiation is reduced compared to clear-sky conditions, since the radiation emanating from clouds is less than from the lower, warmer, levels of the atmosphere which they obscure. On the other hand, clouds at all levels generally have higher albedo than
Symbols for quantities

$\Delta T$  surface air temperature change (K)

$\Delta T_{2\times}$  equilibrium climate sensitivity (K)

$N$  TOA net downward radiative flux (W m$^{-2}$), with components $N_i$

$F$  (net) radiative forcing (W m$^{-2}$) due to $2 \times$ CO$_2$, with components $F_i$

$\alpha$  (net) climate feedback parameter (W m$^{-2}$ K$^{-1}$), $F = \alpha \Delta T$

$Y_i$  component $i$ of climate feedback, positive for feedback which enhances warming (opposite sign convention to $\alpha$)

Subscripts denoting components

LN  clear-sky longwave

SN  clear-sky shortwave

LC  cloud longwave

SC  cloud shortwave

L  net longwave, the sum of LN and LC

S  net shortwave, the sum of SN and SC

N  clear-sky, the sum of LN and SN

C  cloud, the sum of LC and SC

Table 1: Summary of notation

the surface, and increase the TOA outgoing shortwave radiation.

a. HadSM3 model

Using the HadSM3 model, an experiment is carried out in which CO$_2$ is instantaneously quadrupled. We refer to this as a “transient experiment”. HadSM3 comprises the HadAM3 atmosphere GCM (Hadley Centre atmosphere model version 3) coupled to a “slab” ocean i.e. a mixed layer with prescribed horizontal heat convergence (Williams et al., 2001). The version of HadSM3 used here is not identical with that used by Gregory et al. (2004), but our results are consistent with theirs and for the purpose of this work we regard them as the same model, which we call “standard HadSM3”. (The present version includes the sulphur cycle, which was added by Murphy et al. (2004); we have no evidence of significant differences caused by this to our results.)

In the transient experiment, the climate initially evolves rapidly, the rate of change decreasing as the new steady state is approached. Figure 1a shows that changes in longwave/shortwave clear-sky/cloud $N_i$ separately depend linearly on $\Delta T$. (All quantities are differences from a control run with unchanged CO$_2$.) Regressions are made from the first 20 years of data, during which the $4 \times$ CO$_2$ steady state is largely achieved (cf. Gregory et al., 2004). The regression coefficients are shown in Table 2. The forcings $F_i$ have been divided by two to make them apply to $2 \times$ CO$_2$. The uncertainties (shown as one standard error) have been calculated assuming ten degrees of freedom, on the basis of the autocorrelation in the timeseries.

The majority of $F$ (the $N$-intercept) is longwave clear-sky $F_{LN}$, as expected for CO$_2$. As noted above (Section 1), forcings diagnosed in this way include stratospheric adjustment; the instantaneous forcing is larger than the adjusted forcing (e.g. Cess et al., 1993; Myhre et al., 1998). The longwave clear-sky $Y_{LN}$ is the largest $Y_i$ (in magnitude) and includes the black-body, water-vapour and lapse-rate feedbacks (separately quantified by Colman et al., 2001; Colman, 2003). Their sum is negative, and dominates $\sum Y_i$, so that $\alpha > 0$ as required for stability i.e. overall negative feedback. (The sign convention for the $Y_i$ is chosen so that a positive $Y_i$
Figure 1: The evolution of annual-mean global-mean radiative fluxes $N$ at the TOA with annual-mean global-mean surface air temperature change $\Delta T$ in experiments in which CO$_2$ is instantaneously quadrupled using the standard and modified HadSM3 slab models; the results have been divided by two before plotting, to make them applicable to 2 $\times$ CO$_2$. The symbols show annual means and the lines are regressions.
enhances the warming.)

The instantaneous shortwave forcing at the tropopause is negative, because of increased shortwave absorption by CO$_2$ in the stratosphere. Part of the stratospheric adjustment is a compensating increase in longwave radiation, mostly to space. The sum of adjusted shortwave and longwave forcing must be the same at TOA and tropopause because the heat content of the stratosphere changes negligibly after the adjustment is complete, but the partition of the forcing between longwave and shortwave is different at the two levels because of shortwave absorption and longwave emission by the stratosphere. At the top of the atmosphere, the shortwave forcing is positive. However, it is small in any case, being only ~ 4% of the adjusted longwave forcing at the tropopause (Mylehe et al., 1998). The (stratosphere-adjusted) TOA $F_{SN}$ in HadSM3 is ~ 2% of $F_{LN}$. The shortwave clear-sky $Y_{SN}$ is an important positive feedback, caused by the reduction of surface albedo as the area of snow and ice contracts, and increased absorption by water vapour.

The cloud components of forcing are surprising. Though of a smaller magnitude than $F_{LN}$, both $F_{LC}$ and $F_{SC}$ are significantly different from zero at the 5% level; $F_{LC} < 0$ and $F_{SC} > 0$. As these terms have similar sizes, the net $F = 3.3 \pm 0.2$ W m$^{-2}$ is not very different from the clear-sky forcing. The partition of $F$ into longwave $F_L = 2.52 \pm 0.16$ W m$^{-2}$ and shortwave $F_S = 0.78 \pm 0.10$ W m$^{-2}$ is affected, however, $F_S$ being about 25% of $F$, a much larger fraction than in the clear-sky.

It is notable that $N_{SC}$ has relatively large interannual variability, leading to a bigger uncertainty in $F_{SC}$ than in the other $F_i$. A similar point was made by Colman et al. (2001). This suggests that internal variability of the climate system has its largest effect on the radiation budget through fluctuations in low cloud, which has a shortwave but not a large longwave radiative effect. The same suggestion has been made regarding observed TOA variations (Forster and Gregory, 2006).

We interpret the non-zero $F_{LC}$ and $F_{SC}$ as evidence of adjustment of clouds to the extra CO$_2$ in the atmosphere. We suggest that this is caused by rapid local changes to the vertical temperature profile of the atmosphere due to the altered radiative heating, with consequent changes to stability, vertical mixing and the moisture profile, analogous to the changes causing the indirect and semi-direct aerosol forcings. We refer to these induced cloud contributions to $F$ as “tropospheric adjustment” of CO$_2$ forcing. They are like stratospheric adjustment in that they occur in much less than a year, in response to the elevated CO$_2$ concentration. It would be valuable to examine the processes of tropospheric adjustment in more detail, for instance by analysis of the heating rates and fluxes through the depth of the atmosphere, as was done by Sokolov (2006), who also comments on the contribution of tropospheric adjustment to forcing.

The method used to partition fluxes into clear-sky and cloud can give misleading results (Soden et al., 2004). Increasing the CO$_2$ concentration instantaneously reduces atmospheric longwave emission, but has no instantaneous effect on cloud longwave emission, so the instantaneous all-sky longwave forcing of CO$_2$ is smaller than it would be for clear skies. The difference between all-sky and clear-sky contributes to $F_{LC}$. However, this cloud “masking” effect is considerably smaller than our $F_{LC}$. Instantaneous all-sky forcing (sum of longwave and shortwave) is 0.16 W m$^{-2}$ (4%) smaller than clear-sky in the BMRC GCM (Colman et al., 2001). Instantaneous all-sky longwave forcing is on average 14% smaller than clear-sky in the set of GCMs analysed by Cess et al. (1993). In our results for HadSM3, however (Table 2), longwave all-sky (not instantaneous i.e. after tropospheric adjustment) $F_L = F_{LN} + F_{LC}$ is ~30% smaller than $F_{LN}$ (on average 25% in the set of models considered in Section 2b). The non-zero $F_{SC}$ is even less likely than $F_{LC}$ to be a cloud masking effect, since the instantaneous clear-sky shortwave CO$_2$ forcing is much smaller than the longwave, and smaller than $F_{SC}$ itself in magnitude. We
Table 2: Regression of global-mean radiative fluxes $F_i$ at the top of the atmosphere against global-mean surface air temperature $\Delta T$ in experiments with various slab models in which CO$_2$ is instantaneously doubled. $F$ is the intercept on the $N$-axis (the forcing) and $Y$ is the slope (the feedback parameter). (In the HadSM3 experiments, CO$_2$ is quadrupled, and results for $F$ have been divided by two.) The uncertainties for individual models are $\pm 1$ standard errors from the regression. The uncertainties for the ensemble are $\pm 1$ standard deviations. The names of the models from the CMIP3 database are more properly the names of atmosphere–ocean GCMs (AOGCMs) with the same atmosphere component as the slab models \textit{e.g.} "GISS-ER" here refers to the GISS-E atmosphere model coupled to a slab ocean, although actually GISS-ER is an AOGCM which includes GISS-E. The atmosphere model of HadSM3 is the same as that of the UKMO-HadCM3 AOGCM in the CMIP3 database. The atmosphere model of MIROC3.2(LS) is the same as that of the MIROC3.2(medres) AOGCM in the CMIP3 database, but MIROC3.2(HS) is \textit{not} the same as MIROC3.2(hires) in the CMIP3 database. The model names are abbreviated in the second set of results.

<table>
<thead>
<tr>
<th>Model</th>
<th>$F_{LN}$</th>
<th>$F_{SN}$</th>
<th>$F_{LC}$</th>
<th>$F_{SC}$</th>
<th>$F$</th>
<th>$\Delta T_{2x}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>standard HadSM3</td>
<td>3.66 ± 0.08</td>
<td>0.07 ± 0.05</td>
<td>-1.14 ± 0.09</td>
<td>0.71 ± 0.12</td>
<td>3.30 ± 0.17</td>
<td></td>
</tr>
<tr>
<td>modified HadSM3</td>
<td>3.84 ± 0.06</td>
<td>0.15 ± 0.03</td>
<td>-1.09 ± 0.06</td>
<td>-0.27 ± 0.30</td>
<td>2.63 ± 0.32</td>
<td></td>
</tr>
<tr>
<td>CCSM3</td>
<td>3.79 ± 0.08</td>
<td>0.00 ± 0.17</td>
<td>-0.74 ± 0.14</td>
<td>-0.12 ± 0.19</td>
<td>2.93 ± 0.23</td>
<td></td>
</tr>
<tr>
<td>CGCM3.1(T47)</td>
<td>3.56 ± 0.14</td>
<td>0.18 ± 0.20</td>
<td>-0.73 ± 0.15</td>
<td>0.99 ± 0.28</td>
<td>4.00 ± 0.35</td>
<td></td>
</tr>
<tr>
<td>CSIRO-Mk3.0</td>
<td>3.44 ± 0.13</td>
<td>0.06 ± 0.10</td>
<td>-0.88 ± 0.15</td>
<td>0.52 ± 0.33</td>
<td>3.14 ± 0.34</td>
<td></td>
</tr>
<tr>
<td>GISS-ER</td>
<td>4.49 ± 0.13</td>
<td>-0.01 ± 0.14</td>
<td>-1.13 ± 0.26</td>
<td>0.40 ± 0.28</td>
<td>3.75 ± 0.27</td>
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</tr>
<tr>
<td>MIROC3.2(HS)</td>
<td>3.40 ± 0.11</td>
<td>0.18 ± 0.12</td>
<td>-0.56 ± 0.09</td>
<td>0.49 ± 0.23</td>
<td>3.51 ± 0.25</td>
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</tr>
<tr>
<td>MIROC3.2(LS)</td>
<td>3.47 ± 0.15</td>
<td>0.14 ± 0.13</td>
<td>-0.62 ± 0.12</td>
<td>1.02 ± 0.41</td>
<td>4.02 ± 0.45</td>
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</tr>
<tr>
<td>MRI-CGCM2.3.2</td>
<td>3.62 ± 0.21</td>
<td>-0.42 ± 0.13</td>
<td>-0.77 ± 0.25</td>
<td>0.55 ± 0.45</td>
<td>2.98 ± 0.48</td>
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</tr>
<tr>
<td>UKMO-HadGEM1</td>
<td>3.63 ± 0.19</td>
<td>-0.30 ± 0.19</td>
<td>-0.85 ± 0.17</td>
<td>0.57 ± 0.40</td>
<td>3.05 ± 0.52</td>
<td></td>
</tr>
<tr>
<td>ensemble</td>
<td>3.69 ± 0.31</td>
<td>0.01 ± 0.21</td>
<td>-0.85 ± 0.21</td>
<td>0.49 ± 0.41</td>
<td>3.33 ± 0.47</td>
<td></td>
</tr>
</tbody>
</table>

Radiative forcing (W m$^{-2}$)

$F_{LN}$, $F_{SN}$, $F_{LC}$, $F_{SC}$, $F$
therefore think that $F_C = F_{LC} + F_{SC}$ is not an artefact of cloud masking.

The cloud feedbacks $Y_{LC}$ and $Y_{SC}$ are rather weak, especially the latter i.e. the regression lines are almost flat. Most of the changes in cloud are the rapid tropospheric adjustment. The changes in cloud due to changing climate, as measured by $\Delta T$, are relatively small. The usual feedback calculations done from the steady state count all cloud changes as feedback, but this time-dependent calculation indicates that most should be regarded as forcing.

Murphy et al. (2004) and Webb et al. (2006) constructed ensembles of modified versions of HadSM3 by varying parameters, with the aim of quantifying the systematic uncertainty in climate feedbacks. The second ensemble (Webb et al., 2006) varied 26 parameters in combination. From this ensemble we identified one model, which we refer to as “modified HadSM3”, having a particularly low $\Delta T_{2x}$ of 2.5 K. Two $4 \times CO_2$ experiments were done with modified HadSM3, and their results averaged together year by year, in order to reduce the influence of interannual variability. The plot of $N_i$ against $\Delta T$ is shown in Figure 1b and the regression results in Table 2.

There are no statistically significant differences at the 5% level between the corresponding feedback parameters of standard and modified HadSM3. (The uncertainties are generally smaller for the modified model because they are calculated from an ensemble of two experiments.) The reason for the lower $\Delta T_{2x}$ is therefore not different climate feedback, according to our definition, but lies in $F_{SC}$, which is positive $(+0.71 \pm 0.12 \text{ W m}^{-2})$ in the standard HadSM3 and negative $(-0.27 \pm 0.30 \text{ W m}^{-2})$ in the modified model. The difference between the values is significant at the 5% level. Webb et al. (2006) show that variants of HadSM3 that have low $\Delta T_{2x}$ in the ensemble of which our two versions are members generally show increases in the amount and optical thickness of low clouds, which tend to have a negative radiative effect, dominated by shortwave. The other $F_i$ are indistinguishable between the two models. Consequently $CO_2$ has a smaller $F = 2.6 \pm 0.3 \text{ W m}^{-2}$ for $2 \times CO_2$ in the modified HadSM3, significantly less at the 5% level than the standard HadSM3 value.

### b. Intercomparison of climate models

In connection with the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change, a large database has been assembled of results from many GCMs. This database, known as “CMIP3”, includes experiments done with several other slab models under instantaneous doubling of $CO_2$ and their corresponding control experiments. For some slab models, the CMIP3 database unfortunately contains only a sample of years from the steady state with $2 \times CO_2$, omitting the initial period of warming, which is of interest to us here, but which was not included in the protocol for CMIP3 experiments. The Cloud Feedback Model Intercomparison Project (Webb et al., 2006) collected data with this initial period from several slab models, including two versions with different cloud physics of the MIROC3.2 slab model, denoted by “LS” and “HS” for low and high climate sensitivity. The former model is in the CMIP3 database; the latter is described by Ogura et al. (2006).

The variation of $N$ with $\Delta T$ for eight slab models is shown in Figure 2. Results for CGCM3.1 (T63) are very similar to those of the corresponding lower-resolution T47 model shown. Like standard and modified HadSM3, all these models yield good straight lines, indicating time-independent forcing $F$ and climate feedback parameter $\alpha$. This is a notable result, because it shows that $H \propto \Delta T$ in all climate states of these slab models, not only in steady states. However, there is evidence from AOGCMs that climate feedbacks are somewhat dependent on climate state (Senior and Mitchell, 2000; Boer and Yu, 2003a), probably because some aspects of the geographical pattern of surface climate change evolve slowly, on the timescales of
Figure 2: The evolution of annual-mean global-mean radiative fluxes $N$ at the TOA with annual-mean global-mean surface air temperature change $\Delta T$ in experiments with various slab models in which CO$_2$ is instantaneously doubled. Note that the panels all have different axis ranges. The key for the linestyles and symbols is the same as in Figure 1.
ocean climate change, and $N$ vs. $\Delta T$ then deviates from a straight line (Gregory et al., 2004). Such behaviour reopens the issue of distinguishing forcing from feedback on this multicentury timescale, but we do not pursue the question here.

The main qualitative features of the regression coefficients are also common among models (Table 2). The clear-sky longwave components are most similar across models, with coefficients of variation (standard deviation divided by mean) of 6% in $Y_{LN}$ and 8% in $F_{LN}$. In most models, the clear-sky shortwave $F_{SC}$ is not significantly different from zero at the 5% level. The net forcing $F = 3.7 \pm 0.3 \text{ W m}^{-2}$ agrees with the forcings of 3.71 W m$^{-2}$ calculated by Myhre et al. (1998), and $3.7 \pm 0.2 \text{ W m}^{-2}$ reported by Forster and Taylor (2006) for a different set of GCMs.

All the models show non-zero cloud components of forcing, both longwave $F_{LC}$ and shortwave $F_{SC}$, which tend to compensate to some extent, as in standard HadSM3. In all models, $F_{LC}$ is significantly less than zero at the 5% level. In most models, $F_{SC}$ is positive, but it is significantly greater than zero in only a few cases. In all but two models $F_C = F_{LC} + F_{SC}$ is negative, so the net troposphere-adjusted forcing $F < F_N = F_{LN} + F_{SN}$; the exceptions are CGCM3.1(T47) and MIROC3.2(LS), which have unusually large $F_{SC} \approx +1 \text{ W m}^{-2}$. The ensemble-mean $F_C = -0.4 \pm 0.5 \text{ W m}^{-2}$. The mean is significantly less than zero at the 5% level if the 10 models are regarded as independent instances from a common distribution, so that the standard error of the mean is $\sqrt{10} = 3.2$ times smaller than the inter-model standard deviation indicated.

Stowasser et al. (2006) applied the method of Gregory et al. (2004) to three AOGCMs forced by increased insolation. Their results also give evidence of a negative $F_C$ (they do not distinguish longwave and shortwave) which reduces the net solar forcing in the CSM1 and CCSM2 AOGCMs (the stars in their Figure 3 point to a negative intercept on the $N$-axis, which they call $\langle R' \rangle$). The omission of $F_C$ from the forcing could account for the apparent inconstancy of the climate sensitivity and feedback parameters calculated by Stowasser et al. (2006) in these two models (their Figures 1 and 2), whereas they find a more constant $\alpha$ for CGCM3, which has a small $F_C$ (Lambert and Faull, 2007).

Modified HadSM3 has the smallest $\Delta T_{2x} = 2.5 \text{ K}$; its net climate feedback parameter $\alpha$ matches the ensemble mean of about 1.0 W m$^{-2}$ K$^{-1}$, but its negative $F_{SC}$ is unusual. CCSM3 also has $F_{SC} < 0$ and consequently the second lowest $\Delta T_{2x} = 2.7 \text{ K}$, since it also has a typical $\alpha$. The tropospheric adjustment to CO2 thus appears to be a substantial influence on climate sensitivity. Net forcing $F$ has a coefficient of variation of 14%, to which $F_{SC}$ makes the largest contribution.

$Y_{SC}$ is significantly different from zero at the 5% level only in MIROC3.2(HS). That is, with the exception of this one model, the shortwave radiative response to CO2 due to clouds is not a climate feedback; it may depend on the CO2 concentration, but not on global warming. The longwave radiative response due to clouds is a combination of generally negative forcing with generally positive feedback, but this feedback is small compared with $Y_{LN}$ and $Y_{SN}$. From the statistics of the ensemble, $Y_{LC}$ is $-7 \pm 7\%$ of $Y_{LN}$, but $F_{LC}$ is $-23 \pm 6\%$ of $F_{LN}$.

Our analysis for CCSM3 differs from that of Kiehl et al. (2006) for the same slab model (although at different resolution: T85 here and T42 in their paper). They obtain a similar $\Delta T_{2x} = 2.5 \text{ K}$ but a forcing of 3.58 W m$^{-2}$ from regression on an $N$ vs. $\Delta T$ plot (their Figure 1) like ours (our Figure 2a), whereas we obtain $F = 2.93 \pm 0.23 \text{ W m}^{-2}$. On their plot $N \approx 1 \text{ W m}^{-2}$ in the final steady state, but we would expect it to be zero. This imbalance should have been subtracted, shifting the values in their Figure 1 downwards by $\sim 1 \text{ W m}^{-2}$, and reducing their forcing estimate to agree with ours (Kiehl, pers. comm.). This adjustment does not affect the slope and hence the estimate of the climate feedback parameter, for which we both obtain $\alpha = 1.1 \text{ W m}^{-2} \text{K}^{-1}$. It is also notable that the scatter in their plot is larger and the correlation
Radiative forcing (W m$^{-2}$)

$F_{\text{LN}}$ (clear-sky longwave)

$F_{\text{SN}}$ (clear-sky shortwave)

$F_{\text{LC}}$ (cloud longwave)

$F_{\text{SC}}$ (cloud shortwave)

$F$ (net)

Models:

- Standard HadSM3
- Modified HadSM3

Methods:

- Regression
- Fixed SST

Figure 3: Comparison of components of forcing for 2 $\times$ CO$_2$ evaluated using the regression method of Gregory et al. (2004) and the fixed-SST method of Hansen et al. (2002) in two versions of the HadSM3 slab model.

coefficient consequently lower than in ours.

GISS-ER has $F_{\text{LN}}$ considerably larger than any other model’s, and consequently the largest $F = 3.8 \pm 0.3$ W m$^{-2}$. For this model, Hansen et al. (2005) report the same $\Delta T_{2x}$ as us and a forcing of 4.11 W m$^{-2}$ obtained by a different method (see below). The strong forcing is countered by an unusually strong (negative) $Y_{\text{LN}}$, and accompanied by an unusually weak (positive) $Y_{\text{SN}}$, leading to the third smallest $\Delta T_{2x} = 2.7$ K.

MIROC3.2(HS) has the largest $\Delta T_{2x} = 6.3$ K by a substantial margin, due mostly due to its exceptionally strong positive $Y_{\text{SC}}$. Its other forcing and feedback components are typical. UKMO-HadGEM1 has the second largest $\Delta T_{2x} = 4.5$ K, mostly due to its strong $Y_{\text{SN}}$, largely caused by strong sea ice sensitivity to warming (Johns et al., 2006). UKMO-HadGEM1 shares with MRI-CGCM2.3.2 the feature that $F_{\text{SN}} < 0$. In a way analogous to the tropospheric adjustment of clouds, this indicates some change affecting clear-sky shortwave absorption (e.g. in water vapour or surface albedo) which comes about much more rapidly than global climate change in response to CO$_2$. See Section 4 for further comment on this.

The ensemble standard deviation of $\Delta T_{2x}$ is 1.1 K. If we exclude MIROC3.2(HS) and UKMO-HadGEM1, the two models with a high $\Delta T_{2x}$ due to large shortwave feedbacks, the standard deviation is reduced to 0.5 K, which is due roughly equally to uncertainty in $F$ and in $\alpha$ (combined in quadrature). If all the ten models had the ensemble-mean $F_N$ and $\alpha$, the standard deviation of $\Delta T_{2x}$ would be 0.5 K due to the spread in $F_C$ alone.
An independent method to evaluate the forcing is by running an experiment in which the forcing agent is included but climate feedback is inhibited by not allowing surface temperatures to change. The advantage of this approach is that forcing can be evaluated to arbitrary precision by running a long enough experiment. The disadvantages compared with the regression approach are that this special experiment is needed, and that it gives no information about feedbacks.

In such an experiment $\Delta T = 0 \Rightarrow N = F - \alpha \Delta T = F$, so the forcing is diagnosed as the change in net downward TOA radiation. Hansen et al. (2002) proposed the method, which they implemented by prescribing observed sea-surface temperatures and sea-ice; we will call this a “fixed-SST experiment”. Global climate change is not completely prevented since land surface conditions are allowed to evolve. Hansen et al. (2005) corrected for this by evaluating $F = N^* + \alpha \Delta T^*$ (* for fixed-SST experiment, equivalent to their equation 1). This assumes that the climate feedback parameter in a fixed SST experiment is the same as in a transient experiment; this is unlikely to be exactly true, since feedbacks over sea areas have largely been inhibited. Hence $\alpha$ may not be the appropriate feedback parameter for the correction.

Shine et al. (2003) avoided the problem by fixing land surface temperatures as well. This is technically more difficult to do in a GCM because it interferes with the rapid coupling of the atmosphere boundary layer to land surface temperatures and hydrology during the diurnal cycle and synoptic variation. Such interference is believed to be less severe over the sea, which has a greater heat capacity and a constantly moist surface. (However, this perception might partly result from the rather simple representation that most AOGCMs employ for the near-surface vertical structure of the ocean mixed layer.)

Both Shine et al. (2003) and Hansen et al. (2005) use observed surface temperatures. Climate feedbacks are somewhat dependent on climate state (Senior and Mitchell, 2000; Boer and Yu, 2003a; Gregory et al., 2004) and the same may be true for climate forcing. Consequently it might be better to use the model’s own control climatology for surface temperature in such experiments. This should be very similar over the sea, since slab models are calibrated to reproduce observed SST, but could differ over the land. We have not investigated this issue further.

We suppose that the same value of forcing should be obtained by preventing climate feedback in a fixed-SST experiment (this section), as by regressing to the limit of zero climate change in a transient experiment (Section 2). Shine et al. (2003) and Hansen et al. (2005) do not consider components of TOA forcing, but we presume that they can also be diagnosed in fixed-SST

<table>
<thead>
<tr>
<th>Component</th>
<th>Standard HadSM3</th>
<th>Modified HadSM3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>regression</td>
<td>fixed-SST</td>
</tr>
<tr>
<td>clear-sky longwave $F_{LN}$</td>
<td>$3.66 \pm 0.08$</td>
<td>$3.41 \pm 0.03$</td>
</tr>
<tr>
<td>shortwave $F_{SN}$</td>
<td>$0.07 \pm 0.05$</td>
<td>$0.34 \pm 0.03$</td>
</tr>
<tr>
<td>cloud longwave $F_{LC}$</td>
<td>$-1.14 \pm 0.09$</td>
<td>$-0.67 \pm 0.02$</td>
</tr>
<tr>
<td>shortwave $F_{SC}$</td>
<td>$0.71 \pm 0.12$</td>
<td>$0.60 \pm 0.04$</td>
</tr>
<tr>
<td>net $F$</td>
<td>$3.30 \pm 0.17$</td>
<td>$3.69 \pm 0.05$</td>
</tr>
</tbody>
</table>

Table 3: Comparison of components of forcing for $2 \times CO_2$ evaluated using the regression method of Gregory et al. (2004) and the fixed-SST method of Hansen et al. (2002) in two versions of the HadSM3 slab model. The symbols $\simeq$ and $\neq$ indicate whether the methods are statistically consistent.

### 3. Forcing diagnosed by preventing climate feedback
experiments according to $F_i = N_i^*$. These suppositions are tested by carrying out fixed-SST experiments for standard and modified HadSM3. Each experiment comprises two integrations—one with control CO$_2$, one with 2 × CO$_2$—and each $N_i$ is evaluated as the difference between time-means in the two integrations.

Table 3 and Figure 3 compare the forcings in standard and modified HadSM3 using the regression and fixed-SST methods. The results qualitatively confirm the hypothesis that the methods agree, since in either model the difference between the two methods for each of the components is much less than the differences among the four components. However, the fixed-SST method tends to give less negative and more positive forcings, and the differences between the methods are statistically significant at the 5% level in many cases. The correction proposed by Hansen et al. (2005) would further increase the fixed-SST net forcing. The integrations are 20 years long or more, but the uncertainties have been calculated assuming ten degrees of freedom to allow for interannual autocorrelation, as for the regression method.

Nonetheless, the fixed-SST method agrees with the regression method that the net forcing $F$ in modified HadSM3 is significantly less than in standard HadSM3, and that this is mostly due to $F_{SC}$. This agreement supports our conclusion from Section 2 that tropospheric adjustment to CO$_2$ is responsible for the low $\Delta T_{2\times}$ of modified HadSM3.

4. Geographical distribution of forcing

Both the regression method of Section 2 and the fixed-SST method of Section 3 can be used to obtain the geographical distribution of the components of forcing, as well as the global mean. Using the fixed-SST experiment, the TOA radiation fluxes $N_i(x) = F_i(x)$ can be diagnosed from the model as functions of geographical location $x$ (latitude–longitude). Alternatively, we can regress $N_i(x, t)$ from the transient experiment at each point individually against global-mean surface air temperature change $\Delta T(t)$, and thus obtain a latitude–longitude field of the regression intercepts for $\Delta T = 0$. This procedure is analogous to the treatment of global-mean $N_i(t)$ to obtain the global-mean forcing. We interpret the field of intercepts as the geographical distribution $F_i(x)$ of local contributions to the global-mean $F_i$. Note that only the global-mean $\Delta T$ is constrained to be zero by the regression method; there is no constraint on local surface air temperature change (see Section 5).

The components of forcing for standard HadSM3 are compared in Figure 4 (two left-hand columns). Some similarity is evident between the methods. There are substantial areas where the results are not statistically different from zero, especially in the cloud components, which have greater temporal variability. The dominant $F_{LN}$ is positive everywhere, has least geographical variation, and is larger at lower latitudes (Colman et al., 2001; Boer and Yu, 2003b). $F_{SN}$ shows interesting features in the Arctic; both methods indicate significant forcing, but they give opposite signs.

The cloud components $F_{LC}$ and $F_{SC}$ result from tropospheric adjustment to CO$_2$. They have the most pronounced geographical variation, with local values of magnitude $> 10$ W m$^{-2}$ greatly exceeding their global means of magnitude $~1$ W m$^{-2}$. We noted that the global means of $F_{LC}$ and $F_{SC}$ tend to have opposite signs (Table 2), and this anticorrelation is evident geographically as well. In the control climate, high cold cloud causes a positive longwave (greenhouse) effect, which is opposed by a negative cloud shortwave (planetary albedo) effect. Figure 4 suggests that when CO$_2$ is added to the atmosphere, there are rapid changes in cloud, predominantly reductions. This produces opposing $F_{LC}$ and $F_{SC}$ at low latitudes (Webb et al., 2006; Sokolov, 2006), which is particularly clear in the Pacific in the regression method. In mid-latitudes,
Figure 4: Geographical distribution of components of $2 \times \text{CO}_2$ forcing in W m$^{-2}$ in standard HadSM3 (two left-hand columns) and modified HadSM3 (two right-hand columns) evaluated using the regression method of Gregory et al. (2004) and the fixed-SST method of Hansen et al. (2002). Areas where the forcing is not significantly different from zero at the 5% level are uncoloured.
reduction in cloud has a weaker effect in the longwave, because clouds are lower and at a more similar temperature to the surface, but it still has a substantial positive effect in the shortwave through reduction of albedo, shown by both methods across Eurasia and North America.

Components of forcing for modified HadSM3 are compared in Figure 4 (two right-hand columns). $F_{LN}$ and $F_{SN}$ are similar to the standard HadSM3. Both the regression and fixed-SST methods show that the positive areas of $F_{SC}$ in northern mid-latitudes are largely eliminated in the modified model. These differences are consistent with the negative global mean of $F_{SC}$, and hence explain the low $\Delta T_{2\times}$ of the modified model.

The regression method also shows differences in cloud components between the models in the Pacific subtropics: in modified HadSM3 positive areas appear in $F_{LC}$, matched by negative areas in $F_{SC}$. This suggests that tropospheric adjustment produces some increases in cloud at low latitudes in this model.

Fixed-SST experiments are not available from the CMIP3 database, but the regression method can be used to evaluate forcing and feedback geographically from the transient $2 \times CO_2$ experiments. Statistical significance is worse than for our HadSM3 experiments, which use $4 \times CO_2$, and hence show a larger signal. Preliminary inspection of the forcing fields (not shown) indicates that they share some general features with HadSM3. $F_{LN}$ is relatively uniform. $F_{LC}$ and $F_{SC}$ have strong spatial variation, with anticorrelation at low latitudes. $F_{SN}$ exhibits large values at high latitudes, in the regions of the sea ice. In several models this forcing is negative, as was noted in Section 2 for the global mean of MRI-CGCM2.3.2. In that model, the sea ice area initially expands and the surface air temperature falls slightly when $CO_2$ is doubled. Further investigation would be needed to establish a physical explanation for this phenomenon.

5. Surface and tropospheric adjustment

The largest differences between the regression and fixed-SST estimates of components of forcing in Table 3 and Figure 3 are in $F_{LC}$. Figure 4 suggests that this may arise from differences in the low-latitude Pacific, where the regression method shows pronounced $F_{LC}$, predominantly negative in the standard HadSM3 and negative in large regions in the modified HadSM3, while the fixed-SST method has much weaker features. By construction, both methods evaluate $F_i = N_i$ for $\Delta T = 0$, but they differ in how they impose $\Delta T = 0$. In the fixed-SST method, the local change must be zero everywhere over the sea, which is the majority of the world, but this is not required by the regression method. For example, for the standard HadSM3, regression indicates mean temperature change over land and sea respectively of $+0.40 \pm 0.05$ K and $-0.16 \pm 0.02$ K for global $\Delta T = 0$. This suggests the idea that the rapid adjustment to $CO_2$ in a transient experiment may include changes in surface temperatures as well as in the troposphere. Tropospheric adjustment in a fixed-SST experiment might therefore be somewhat different because surface temperature change is largely inhibited.

Radiative forcing at the surface caused instantaneously by doubling of $CO_2$ is much smaller than at the TOA (Collins et al., 2006), because the atmosphere is fairly opaque in the longwave. However, a continued difference between the surface and the TOA implies storage of heat in the atmosphere, which has a relatively small heat capacity. On a fairly short timescale of months, during which tropospheric and stratospheric adjustment takes place, this difference must be eliminated. Unlike the instantaneous forcing, the adjusted forcing is therefore identical at the surface, tropopause and TOA (Shine et al., 2003; Hansen et al., 2005).

The adjustment of the net downward surface heat flux may involve changes to non-radiative fluxes, which can be evaluated like those at the TOA by regression against $\Delta T$. Figure 5
Figure 5: The evolution of annual-mean global-mean surface fluxes with annual-mean global-mean surface air temperature change in an experiment with the standard HadSM3 model in which CO$_2$ is instantaneously quadrupled. The results have been divided by two so that they apply to 2 \times CO$_2$. All fluxes are positive downward. Negative downward latent heat implies positive evaporation.

indicates that in HadSM3 the largest part of surface forcing due to 2 \times CO$_2$ is a reduction in latent heat loss due to evaporation, with smaller contributions from increased net downward longwave and shortwave radiation, the latter being consistent with a reduction of cloud amount as an aspect of tropospheric adjustment. As the climate responds to the forcing by warming up, the net downward longwave flux increases, because the atmosphere becomes warmer and moister. Net heat loss by the surface is achieved by a strong increase in evaporative cooling. The change in surface shortwave is small, consistent with weak cloud feedback. We note that global-mean changes in surface heat fluxes are dominated by changes over the sea, which covers most of the world, and changes over land have a different character as evaporation is limited by water availability. Since the surface fluxes are related to the near-surface tropospheric temperature profile, it would not be surprising if they were different in transient and fixed-SST experiments. A similar point is made by Lambert and Faull (2007) when comparing transient and fixed-SST experiments with increased insolation.

The basic distinction made in Section 1 is between the rapid local response to the CO$_2$, which we include in “forcing”, and the slow widespread response over a timescale set by the thermal inertia of the climate system, with a magnitude determined by “feedback”. An initial rapid response (for instance over continental interiors) might change the global-mean surface air temperature as well as the global-mean surface fluxes. However, the regression method could not then be used to evaluate it, because it defines $\Delta T = 0$ as indicating “no climate change”. The extension of the idea of tropospheric adjustment to include changes in the surface temperature and fluxes is an obvious extrapolation, but it needs careful consideration to understand what
it means and whether it is useful.

6. Summary and conclusions

Gregory et al. (2004) proposed a method for diagnosing radiative forcing and climate feedback from transient climate-change experiments, by regressing TOA radiative fluxes against global-mean surface air temperature $\Delta T$. We have demonstrated that this method works in several GCMs for experiments with increased atmospheric CO$_2$ concentrations, and we have extended it to separate the longwave/shortwave clear-sky/cloud contributions to forcing and feedback. The clear-sky forcing from this method gives reasonable agreement with radiative forcing calculations that allow for stratospheric adjustment.

The method suggests a practical distinction between forcing and feedback to be made on the basis of timescale, namely that we should regard all radiative changes as forcing that come about rapidly compared with the timescale of climate change as measured by $\Delta T$, for instance within much less than a year. Our definition of forcing due to CO$_2$ includes not only its instantaneous greenhouse effect and stratospheric adjustment, but any other rapid adjustment of the system which may occur, in particular due to clouds. It is a familiar idea that the radiative effect of cloud changes may be counted as forcing, since this is the case for the indirect and semi-direct effects of aerosols. In all the GCMs examined in the present work, CO$_2$ also is found to induce small but statistically significant cloud forcing. Longwave is usually negative and shortwave usually positive, so they compensate to some degree, but the sum is negative in most models. In some models the induced shortwave forcing is negative, leading to a substantially reduced net forcing by CO$_2$ and a low $\Delta T_{2x}$. Extending the regression method to examine geographical variation provides further evidence for cloud changes.

We regard these cloud changes as a tropospheric adjustment of the forcing due to CO$_2$. The usual interpretation would be to call them cloud feedbacks. We show, on the contrary, that shortwave cloud feedback is statistically insignificant in nearly all the models examined here, and radiative change caused by clouds depends on CO$_2$ change rather than on the magnitude of global climate change (measured by $\Delta T$). Further studies would be useful of how the differences in cloud parametrisations and simulated climates among models lead to different tropospheric adjustments.

We propose that the net forcing including both tropospheric and stratospheric adjustment could be called “effective” forcing, to distinguish it from “instantaneous” forcing. We have shown that this definition of forcing agrees fairly well with results from the method introduced by Hansen et al. (2002), and developed by Shine et al. (2003) and Hansen et al. (2005), of preventing climate feedback by fixing surface conditions in an experiment with the forcing agent included, in which case forcing can be diagnosed as the change in TOA radiative flux. This agreement of methods is reasonable, because in different ways they are both estimating the forcing as the change in TOA radiative flux when no global-mean temperature response to the forcing agent has occurred. A more detailed comparison of the methods is required to establish the differences. Since the surface temperatures are locally fixed in the method of Shine et al. and Hansen et al., the tropospheric adjustment, including changes in surface heat fluxes, may not be the same as implied by the regression method.

On account of tropospheric adjustment, the model spread in $\Delta T_{2x}$ may be partly due to a spread of CO$_2$ forcing, rather than entirely due to differences in the climate feedback parameter. Although this reattribution makes no difference to prediction of steady-state climate change, a correct separation of forcing and feedback is necessary for predicting time-dependent climate
change. As shown in Section 1, different pathways result from different choices of \((F, \alpha)\) that have the same \(\Delta T_{2x} = F/\alpha\); this is relevant to the use of simple climate models to emulate AOGCMs, for example. The possibility of tropospheric adjustment also affects interpretation of past change. If clouds respond directly to tropospheric adjustment due to CO\(_2\), they will “keep up” with rising concentrations, rather than lagging behind with global temperature change; cloud changes in recent decades may be part of the CO\(_2\) forcing rather than a response to it.

Because of these implications, the correct distinction of forcing and feedback is not purely theoretical. If tropospheric adjustment is a real effect, it complicates the definition of these concepts. Rather than seeing this as an issue of terminology, let us concentrate on what is practically important. There are two main points. Firstly, the distinction is most useful if it is made in such a way that the feedback parameter is the same for all kinds of forcing. Recent work (Shine et al., 2003; Hansen et al., 2005) suggests that including non-instantaneous effects in the forcing does have that effect. Secondly, if we include non-instantaneous effects, the distinction is bound to become rather blurred, because there is no longer a clean separation into two timescales which we can obviously designate as forcing and response. The “adjustments” probably occur on various timescales. Our conclusion is therefore that analysis of the response to forcing should not be limited to the final steady states. Examination of time-dependent changes could provide further insight into the systematic uncertainty in model projections of climate change.

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