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Does Model Sensitivity to Changes in CO_2 Provide a Measure of Sensitivity to the Forcing of Different Nature?

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Abstract

Simulation of both the climate of the 20th century and of possible future climate change requires taking into account numerous forcings of different nature. Climate sensitivities of existing general circulation models, defined as the equilibrium surface warming due to increase in atmospheric CO_2 concentrations, vary over a rather wide range. A large number of simulations with the MIT climate model of intermediate complexity with forcings of different nature have been carried out to study to what extent sensitivity to changes in CO_2 concentration represent sensitivities to other forcings. Sensitivity of the MIT model can be changed by changing the strength of the cloud feedback.

Simulations with the versions of the model with different sensitivities show that the sensitivity to changes in CO_2 concentration provides a reasonably good measure of the model sensitivity to other forcings with similar vertical stratifications. However the range of models' responses to the forcings leading to the cooling of the surface is narrower than the range of models' responses to the forcings leading to warming. This is explained by the cloud feedback being less efficient in the case of increasing sea ice extent and snow cover. The range of models' responses to the forcings with different vertical structure, such as increase in black carbon concentration, is also smaller than that for changes in CO_2 concentration.

Contents

1. INTRODUCTION

The MIT 2D climate model has been used in a number of climate change related studies in recent years. Forest *et al*. (2002) used the model to obtain a probability distribution for climate sensitivity consistent with the climate record for the 20th century. This distribution then has been used by Webster *et al*. (2003) for studying uncertainty in future climate change. In both cases, a number of different forcings were considered. In simulations performed by Forest *et al*. (2002), the model was forced by changes in CO_2, sulfate aerosol and ozone. In an ongoing study (Forest *et al.*, 2005) changes in solar constant, volcanic aerosol and vegetation cover are also included. In projections of future climate, changes in different greenhouse gases, ozone, sulfate aerosol and black carbon are taken into account. Climate sensitivities of different versions of the MIT climate model were, however, defined based exclusively on changes in CO_2 concentration. Therefore, it is important to evaluate to what extent model sensitivity to changes in CO_2 characterizes sensitivity to other forcings.

The dependency of the climate system response to external forcing on the nature of the forcing has been a subject of a number of recent studies (*e.g.*, Forster *et al*., 2000; Cook and Highwood, 2004; Hansen *et al*., 1997; Ramaswamy and Chen, 1997). It was shown that the change in surface air temperature, ΔTs, in response to changes in atmospheric CO_2 concentration, solar constant or surface albedo (and some others) is proportional to the adjusted radiative forcing at the tropopause, Fa, regardless of the nature of the forcing, or:

$$\Delta Ts = \lambda Fa, \quad (1)$$

where λ is a climate sensitivity. This is not, however, the case for forcings with significantly different vertical structures such as forcings associated with changes in the concentration of ozone or absorbing aerosols (*e.g.*, Hansen *et al*., 1997). This poses the question of whether model sensitivity to changes in CO_2 concentration characterizes model sensitivities to different forcings, especially to forcings of the second kind.

A number of the equilibrium climate change simulations with versions of the MIT model with different climate sensitivities for a variety of forcings have been performed for this study. A brief description of the model is given in Section 2. Dependence of the model response on the vertical stratification of forcing is discussed in Section 3. In Section 4 results of the simulations with the versions of the model with different sensitivities are discussed. Conclusions are given in Section 5.

2. MODEL DESCRIPTION

The MIT 2D atmospheric model (Sokolov and Stone, 1998) is a zonally averaged statistical-dynamical model developed from the GISS GCM Model II (Hansen *et al.*, 1983). The model includes parameterizations of all the main atmospheric physical processes as well as parameterizations of heat, moisture and momentum transports by eddies. The version used in this study has a latitudinal resolution of 7.8 degrees and 9 vertical layers. Each cell can contain up to four different surface types: land, land ice, ice-free ocean and ocean ice. The model calculates surface temperature, surface and radiative fluxes, and their derivatives with respect to surface temperature separately for different surface types.

A zonally averaged mixed layer model was used as an ocean component in the previous version of the MIT climate model. In the version used in this study, the atmospheric model is coupled to a mixed layer ocean model with a horizontal resolution of 7.8˚ in latitude and 10˚ in longitude. The mixed layer depth is prescribed based on observations as a function of time and location.

The heat flux felt by the ocean model at the point (i, j) is calculated as:

$$F_H(i,j) = F_{HZ}(j) + \frac{\partial F_{HZ}}{\partial T}(j)(Ts(i,j) - Tsz(j)), \quad (2)$$

where $F_{HZ}(j)$ and $\partial F_{HZ}/\partial T(j)$ are zonally averaged heat flux and its derivative with respect to surface temperature; $Ts(i,j)$ and $Tsz(j)$ are the surface temperature and its zonal mean.
The mixed layer model also uses parameterized vertically averaged horizontal oceanic heat transport, the so-called Q-flux. This flux has been calculated from the simulation in which sea surface temperature and sea ice distribution were relaxed to their present-day climatology.

As was shown by Sokolov and Stone (1998), the MIT climate model simulates reasonably well the zonally averaged features of the present-day atmospheric circulation. Both equilibrium and transient responses to an increase in CO_2 concentration produced by the model are similar, in terms of global averaged values and zonal distributions, to the responses obtained in simulations with the 3D general circulation models (GCMs).

3. MODEL RESPONSE TO DIFFERENT FORCINGS, DEPENDENCY ON VERTICAL STRUCTURE OF THE FORCING

A number of 150 year duration equilibrium simulations with the MIT climate model with different forcing have been carried out (**Table 1**). During the last 20 years of the simulations, data required for the radiation calculation have been saved and then used to calculate changes in radiation fluxes and climate feedbacks associated with changes in different climate variables, namely surface temperature, lapse rate, water vapor, cloud cover and surface albedo. Feedbacks were calculated following procedure proposed by Wetherald and Manabe (1988).

To evaluate the model response to changes in black carbon (BC) concentration, an equilibrium climate change simulation (10BC) has been carried out, using changes in black carbon loading simulated by the MIT climate-chemistry model (Wang *et al.*, 1998). Projected changes in BC are rather small and so are forcings associated with these changes. To obtain a statistically significant response changes in BC loading were multiplied by 10. Such an increase in BC leads to a positive forcing of 2.4 W/m^2 at the tropopause and strong negative forcing of –4.0 W/m^2 at the surface (**Figure 1**). In spite of such a strong cooling at the surface, surface temperature actually increases by 1.76K in this simulation. This is explained by the vertical distribution of the BC forcing.

As was shown by Hansen *et al.* (1997), the effectiveness of forcing with respect to surface warming depends on the altitude at which the forcing is applied. They carried out simulations applying a forcing of 4 W/m^2 at each layer of the model and at the surface. Results of those

Table 1. Design parameters for the MIT climate model simulations used in this study.

Simulation	Type of forcing	Forcing at the tropopaus (W/m^2)	Forcing at the surface (W/m^2)	Surface warming (K)	λ K/(W/m^2)
2xCO2	Doubled-CO_2 concentration	3.76	0.8	2.18	0.58
0.5xCO2	Halved-CO_2 concentration	–3.76	–0.56	–2.14	0.57
2%S0	2% increase in solar constant	4.72	3.56	2.28	0.48
–2%S0	2% decrease in solar constant	–4.72	–3.56	–2.22	0.47
ALB	Increase in surface albedo	–3.39	–3.85	–1.54	0.45
STRAER	Increase in stratospheric aerosol concentration	–3.92	–4.08	–1.87	0.48
10BC	Change in black carbon simulated by the MIT climate-chemistry model multiplied by 10	2.36	–4.44	1.76	0.75
LWBC	Fixed longwave forcing with vertical structure of the global and annual mean black carbon forcing	2.36	–4.44	1.76	0.74
LWBCL3	Fixed longwave forcing with the same changes at the top of the atmosphere (TOA) and at the surface as in LWBC, but with “absorbing layer” shifted to 800 hPa	2.36	–4.44	1.23	0.52
LWBCL6	Fixed longwave forcing with the same changes at the TOA and at the surface as in LWBC, but with “absorbing layer” shifted to 320 hPa	2.36	–4.44	0.64	0.27

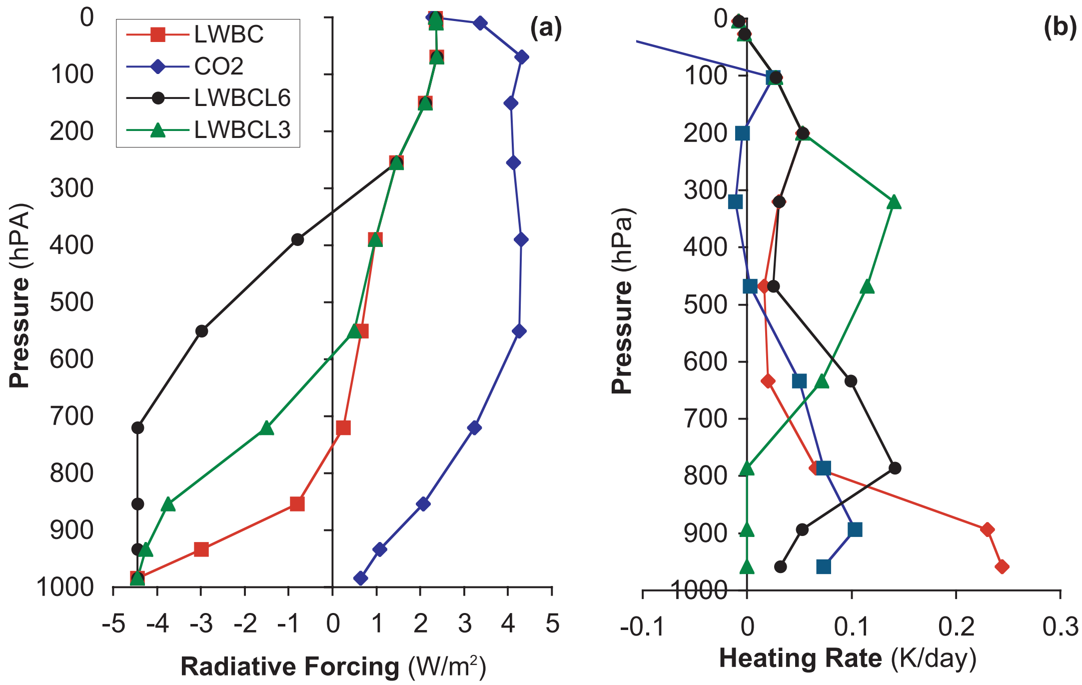

Figure 1. Vertical distribution of (a) radiative fluxes and (b) heating rates in simulations with doubled CO_2 and simulations with "black-carbon-like" ("BC-like") forcing.

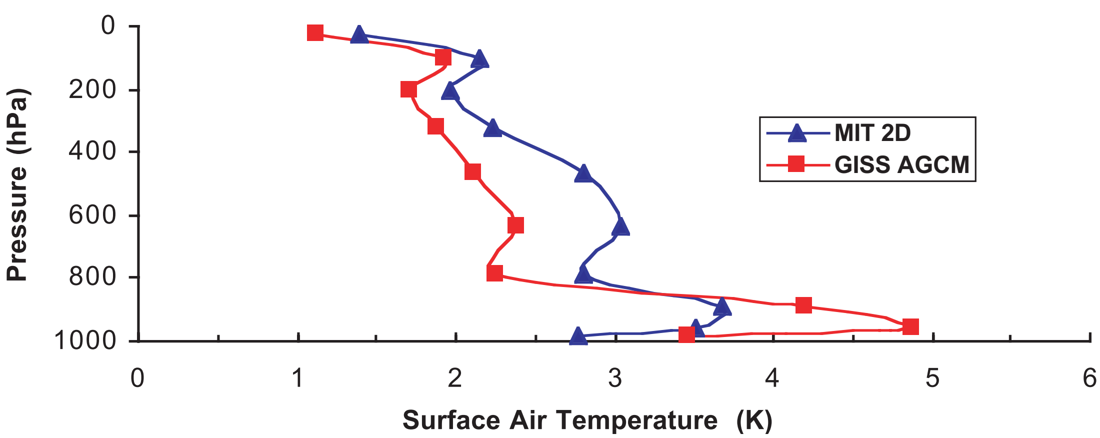

Figure 2. Dependence of surface air temperature increase on the altitude at which forcing was applied, as simulated by the MIT 2D model (triangles), and the GISS AGCM (squares).

simulations are shown in **Figure 2** together with the results of analogous simulations with the MIT 2D model. Both models show a similar dependency, but the GISS model is noticeably more sensitive to the forcings applied in the low troposphere and at the surface. The two top layers (8 and 9) and a part of layer 7 are located in the stratosphere and, as indicated by Hansen *et al.* (1977), when the forcing is applied in those layers an adjusted forcing on the tropopause is significantly smaller than the applied forcing. Therefore, only results for simulations with forcings in the 6 lowest layers are discussed below. The two feedbacks that show the largest differences are lapse rate feedback and cloud feedback (**Table 2**). The lapse rate feedback is negative in all simulations. It is weakest for the forcing in layer two and becomes much stronger when the forcing is applied in the top layers. These differences to a large extent are offset by

Table 2. Strengths of different feedbacks in simulations with 4 W/m^2 forcing applied at different heights (lapse rate, LR; water vapor, Q; clouds, CL; surface albedo, ALB).

L	Height (hPa)	LR	Q	LR+Q	CL	ALB	L+Q+C+A
0	984	−0.34229	1.50329	1.16100	−0.13143	0.35204	1.38161
1	958	−0.27586	1.48009	1.20423	0.23215	0.30544	1.74183
2	894	−0.14481	1.44507	1.30026	0.37037	0.28929	1.95993
3	786	−0.29161	1.48904	1.19743	−0.08282	0.32935	1.44397
4	633	−0.42274	1.55814	1.13540	−0.14018	0.33824	1.33346
5	468	−0.61640	1.70413	1.08773	−0.28229	0.34565	1.15109
6	320	−0.97663	2.00795	1.03132	−0.72512	0.37552	0.68172

changes in the water vapor feedback, but the sum of these two feedbacks is still 30% larger for the forcing in the second layer than for the forcing in layer 6. The cloud feedback is positive when the forcing is applied in the two lowest layers and also becomes strongly negative when the forcing is applied in the top layers.

Cloud cover decreases in all simulations. Surface warming leads to the decrease in high clouds. The decrease in high clouds, due to their relatively small albedo and large difference between temperature of the cloud top and surface temperature, leads to the decrease in the net radiative flux at the top of the atmosphere (TOA) and to the negative cloud feedback. Warming of a particular model layer causes an additional decrease in cloud cover in that layer. Thus, a decrease of low clouds, caused by forcing at the second layer, leads to positive cloud feedback.

The feedbacks shown in Table 1 and later in the paper are calculated as changes in the radiation balance at the tropopause associated with changes in a particular climate variable, such as clouds or water vapor, divided by the change in surface air temperature caused by the particular forcing. These were calculated following the procedure proposed by Wetherald and Manabe (1988). We note that in some studies (*e.g.*, Hansen *et al.*, 1984; Schlesinger and Mitchell, 1987) feedbacks are normalized by surface air temperature feedback.

Hansen *et al.* (1997) refer to cloud feedback associated with absorbing aerosol as the "semi-direct" aerosol effect. Changes related to the tropospheric warming can be separated from changes caused by surface warming in a number of ways. For example, by performing parallel simulations with fixed surface temperature (Cook and Highwood, 2004). It also can be done from the results shown above assuming that changes caused by warming of a particular atmospheric layer and changes caused by surface warming are additive and that the latter are proportional to surface temperature increase.

In **Table 3** the first three columns show changes in surface temperature and radiation fluxes at the tropopause due to changes in lapse rate and water vapor (together) and clouds. "Semi-direct" forcings, shown in columns four and five, are calculated in the following way:

$$F_X^L = H_X^L - \frac{H_X^0}{\Delta T_S^0}\Delta T_S^L \qquad (3)$$

where H_X^L is the change in radiation flux due to the change in variable X (CLD or LR+Q) in the simulation with forcing applied at a layer L and ΔT_S^L is the change in surface air temperature

Table 3. Changes in radiation fluxes at the tropopause due to changes in clouds and lapse rate and water vapor and their fractions not related to surface warming.

ΔTs	H_{LR+Q}	H_{CLD}	F_{LR+Q}	F_{CLD}	F_{SMD}	F_{COR}	λ_{COR}
2.08	2.26584	–0.3253	0	0	0	4	0.52
2.77	3.11389	0.58923	0.0964	1.02244	1.11884	5.11884	0.54
3.02	3.67666	1.04652	0.38683	1.51883	1.90567	5.90567	0.51
2.24	2.52665	–0.27084	0.08651	0.07948	0.166	4.166	0.54
2.02	2.16239	–0.35792	–0.03809	–0.042	–0.08009	3.91991	0.52
1.92	2.01108	–0.63741	–0.08046	–0.33713	–0.4176	3.5824	0.54
1.55	1.59897	–1.23355	–0.08952	–0.99114	–1.08066	2.91934	0.53

in the same simulation. The "corrected" forcing is calculated as a sum of adjusted and total "semi-direct" forcings and the model sensitivity, λ, is calculated from Equation 1. As can be seen, sensitivity defined in such a way shows practically no dependency on forcing and is very close to values obtained in simulations with changes in CO_2, solar constant or stratospheric aerosol load.

The change in radiation fluxes caused by an increase in the loading of black carbon leads to warming concentrated in the two lowest model layers with a maximum around 950 hPa (see Figure 1b). This warming causes a positive feedback that is strong enough to overcome direct cooling at the surface. As was shown by Hansen *et al.* (1997) and Cook and Highwood (2004), the climate impact of the increase in BC concentration strongly depends on the vertical distribution of black carbon.

To evaluate a dependency of the MIT model response to the vertical stratification of the "black-carbon-like" ("BC-like") forcing, three simulations with long wave (LW) forcing have been carried out. In the first simulation (LWBC), LW forcing with the same vertical distribution as an annual mean global mean forcing due to changes in BC has been used. Despite differences in the nature as well as in spatial and temporal patterns of the forcing between simulations 10BC and LWBC, global mean annual responses are very similar in these simulations. Forcings applied in the other two "BC-like" simulations (LWBCL3 and LWBCL6) have the same change at the tropopause and at the surface but the "absorbing layer" is shifted up (see Figure 1). As a result the maximum changes in the heating are concentrated at about 800 and 320 hPa. These forcings lead to a noticeably smaller surface warming, namely 1.23K and 0.64K instead of 1.76K.

The decrease in low clouds in the LWBC simulation (**Figure 3**), while smaller in magnitude than the decrease in high clouds, is nevertheless strong enough to produce positive cloud feedback (**Table 4**). The change in air temperature in the LWBC simulation (Figure 3b), while smaller than in the simulation with forcing applied directly at the surface (SRF), has a similar shape throughout most of the troposphere. In the other two simulations, especially in LWBCL6 simulation, the warming increases faster with height. Differences in the vertical structure of the forcing also affect changes in the hydrological cycle. Changes in both relative humidity and heating due to moist convection show a strong dependence on the vertical structure of the forcing (**Figure 4**). Those differences are reflected in the strengths of different feedbacks (Table 3). Such strong negative lapse rate and cloud feedbacks in a simulation with the forcing concentrated in the upper troposphere lead to the total feedback being negative.

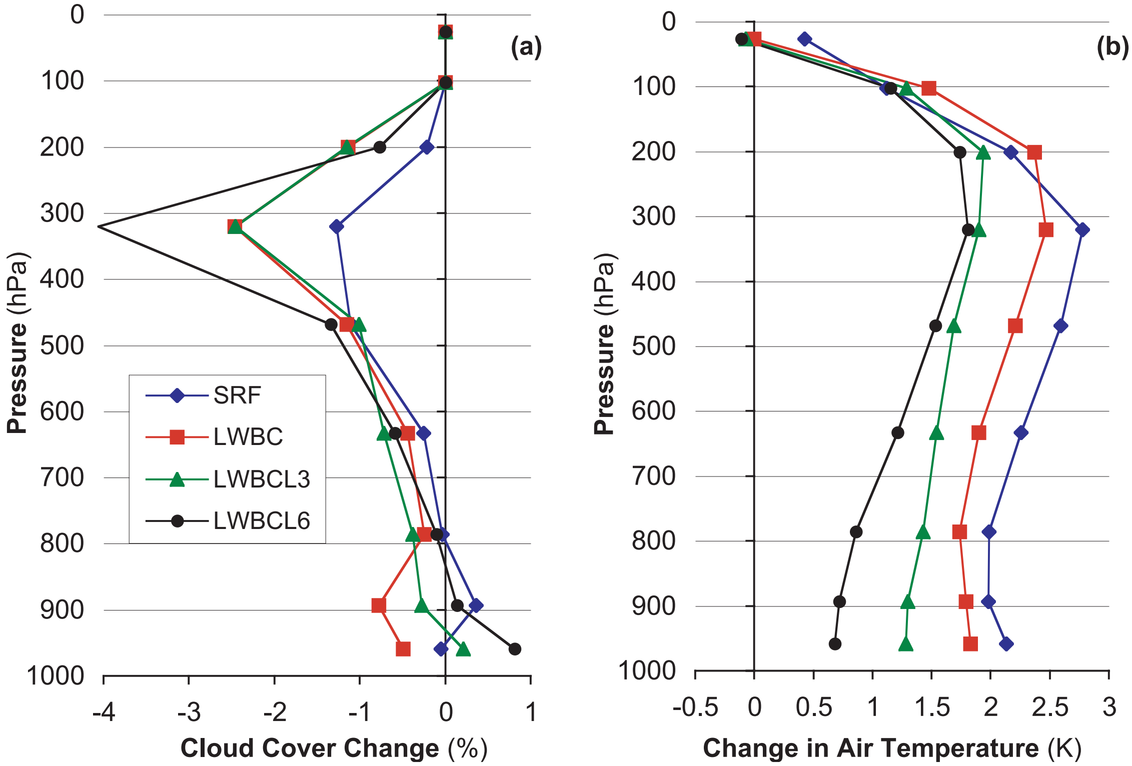

Figure 3. Changes in (a) clouds and (b) air temperature in simulations with surface and “BC-like” forcings.

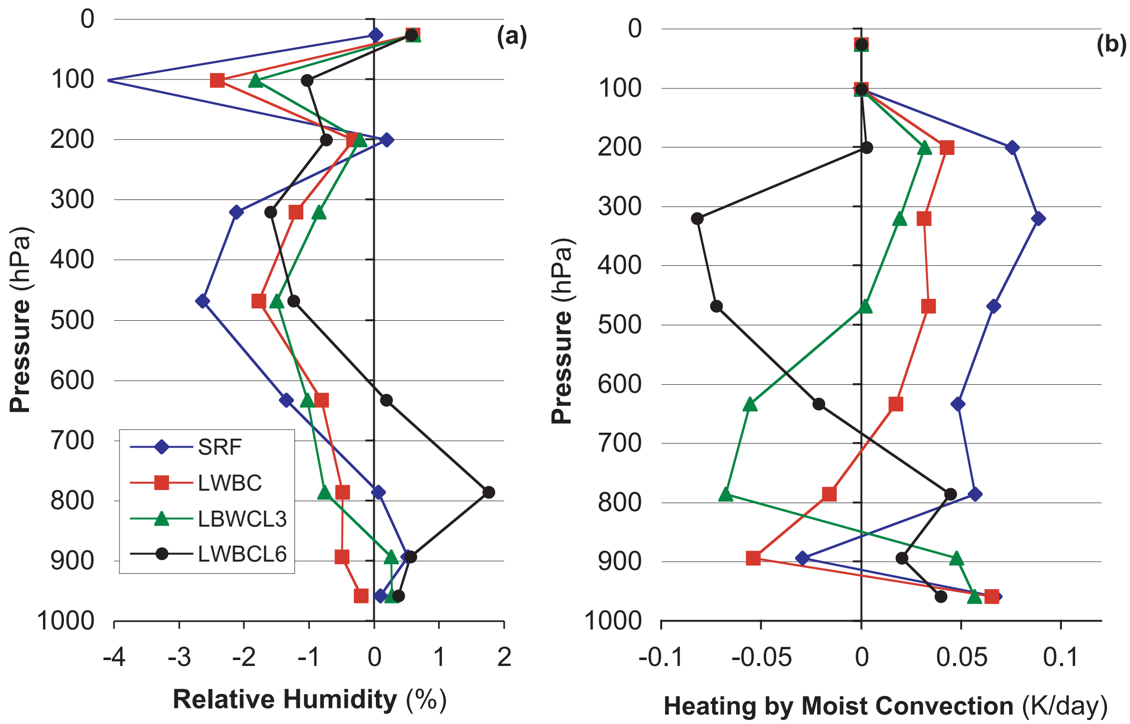

Figure 4. Changes in (a) relative humidity and (b) heating rate due to moist convection in simulations with surface and “BC-like” forcings.

Table 4. Strengths of different feedbacks in simulations with SRF and "BC-like" forcings.

Forcing	LR	Q	LR+Q	CL	ALB	L+Q+C+A
LWBC	–0.405	1.630	1.225	0.361	0.269	1.855
LWBCL3	–0.719	1.789	1.070	–0.134	0.294	1.230
LWBCL6	–2.253	1.827	0.574	–1.254	0.325	–0.356

Table 5. The same as Table 3, but for simulations LWBC, LWBCL3 and LWBCL6.

ΔTs	H_{LR+Q}	H_{CLD}	F_{LR+Q}	F_{CLD}	F_{SMD}	F_{COR}	λ_{COR}
1.78	2.48891	0.54926	0.55181	0.912128	1.463938	3.833938	0.464275
1.23	1.74422	–0.27575	0.405662	–0.025	0.380658	2.750658	0.447166
0.71	0.99252	–1.03182	0.219856	–0.88708	–0.66722	1.702776	0.416966

"Semi-direct" forcings in the last three simulations are shown in **Table 5**. As can be seen from the table, if the components of the cloud feedback and the combined lapse rate and water vapor feedbacks that are not related to surface warming are treated as forcing, Equation 1 provides a good estimate of model sensitivity.

Comparison of the changes in precipitation show that while large-scale precipitation increases in all simulation, convective precipitation decreases in both LWBCL3 and especially in LWBCL6 (**Figure 5**). Changes in precipitation not related to increases in surface temperature were calculated using an equation similar to Equation 3. These changes are negative for all forcings (**Table 6**). Results for simulations with doubled CO_2 are consistent with the findings of Sugi and Yoshimura (2004), who showed that precipitation decreases in simulations with increased CO_2 concentration and fixed sea surface temperature.

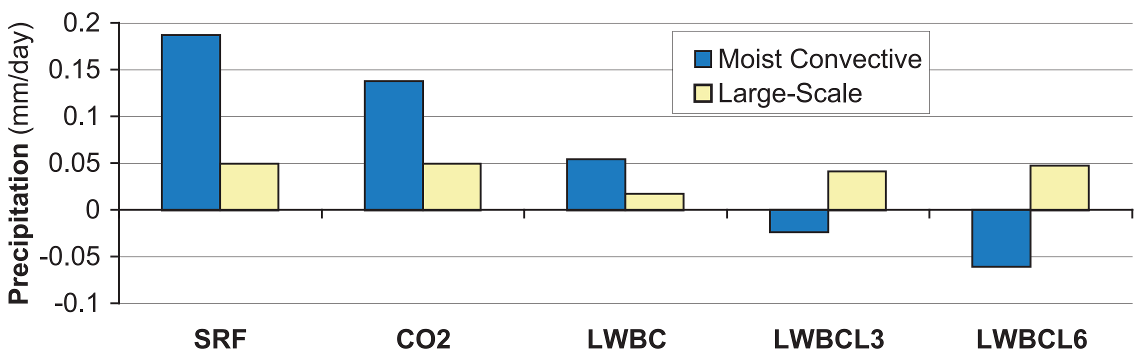

Figure 5. Changes in convective and large scale precipitation in simulations SRF, CO2 and three "BC-like" simulations.

Table 6. Change in total precipitation and the fraction not related to surface warming.

Forcing	**Ts** (K)	**Pr** (mm/day)	**dPr/Prc** (%)	**Pr** (mm/day)	**dPr/Prc** (%)
SRF	2.08	0.236	8.26	0	0
CO2	2.18	0.187	6.54	–0.065	–2.28
LWBC	1.78	0.072	2.52	–0.133	–4.66
LWBCL3	1.23	0.018	0.63	–0.124	–4.34
LWBCL6	0.71	–0.013	–0.46	–0.095	–3.34

4. CHANGING SENSITIVITY OF THE MIT MODEL

The sensitivity of the MIT climate model is varied by changing the strength of the cloud feedback (Hansen *et al.*, 1993; Sokolov and Stone, 1998). Namely, the cloud fractions used in the radiation calculation are calculated as:

$$C = C^0 \cdot (1.0 + k \cdot \Delta T_{srf}), \tag{3}$$

where C^0 is the cloud cover calculated by the model, and ΔT_{srf} is the difference in global mean surface air temperature from its value in a control climate simulation. By changing the parameter k, different sensitivities are obtained. For example, with k equal to 0.04 and –0.03, the sensitivities to the CO_2 doubling are 1.4 and 4.1K, respectively. The natural sensitivity of the model (that is, for k = 0) is 2.2K.

Use of Equation 3, however, leads to the simultaneous increase/decrease in both high and low clouds, which, as mentioned above, have effects of opposite sign on climate sensitivity. As such, if Equation 3 with k = –0.03 is applied to low and middle clouds only, the model sensitivity increases up to 5.0K. On the other hand, if only high clouds are changed then the sensitivity decreases to 1.9K. And finally, if k = –0.03 is used for low and middle clouds and k = 0.03 for high clouds, the sensitivity of the model becomes 6.9K.

Changing high and low clouds in opposite directions allows for obtaining the same sensitivity with smaller values of k compared to using the same value of k for all clouds. As such, for sensitivity of 4.1K the value of k = –0.02, instead of –0.03, is required. It decreases the artificial changes in cloud cover in simulations with different sensitivities.

Figure 6 shows a comparison of the results from the doubled-CO_2 equilibrium simulations using the versions of the MIT model with different sensitivities with the results obtained in similar simulations with different GCMs (Meleshko *et al*., 1999; Senior and Mitchell, 1993; Washington and Meehl, 1993; Yao and Del Genio, 1999). Results from the simulations in which the sensitivity of the MIT model was changed using the same value of parameter k for all clouds are shown by diamonds. Triangles indicate result from the simulations in which k of opposite signs were used for high and low clouds. The results of GCMs are shown by squares. Overall the latter way of varying sensitivity of the MIT model produces better agreement with GCMs and was used in the simulation discussed below.

The strengths of different feedbacks for four versions of the MIT model are shown in **Table 7**. Not surprisingly, changes in sensitivity are mainly associated with differences in cloud feedback. Changes in the lapse rate feedback to large extent are compensated by changes in the water vapor feedback. Such compensation between lapse rate and water vapor feedbacks in the doubled-CO_2 simulations is a feature shown by practically all models (*e.g*., see Colman, 2003).

Table 7. Feedbacks in doubled-CO_2 simulations with different climate sensitivities.

ΔTeq	LR	Q	LR+Q	CL	ALB
1.39	–0.052	1.365	1.313	–1.030	0.334
2.18	–0.195	1.488	1.293	0.068	0.258
4.50	–0.289	1.577	1.288	0.959	0.241
7.45	–0.308	1.633	1.325	1.208	0.311

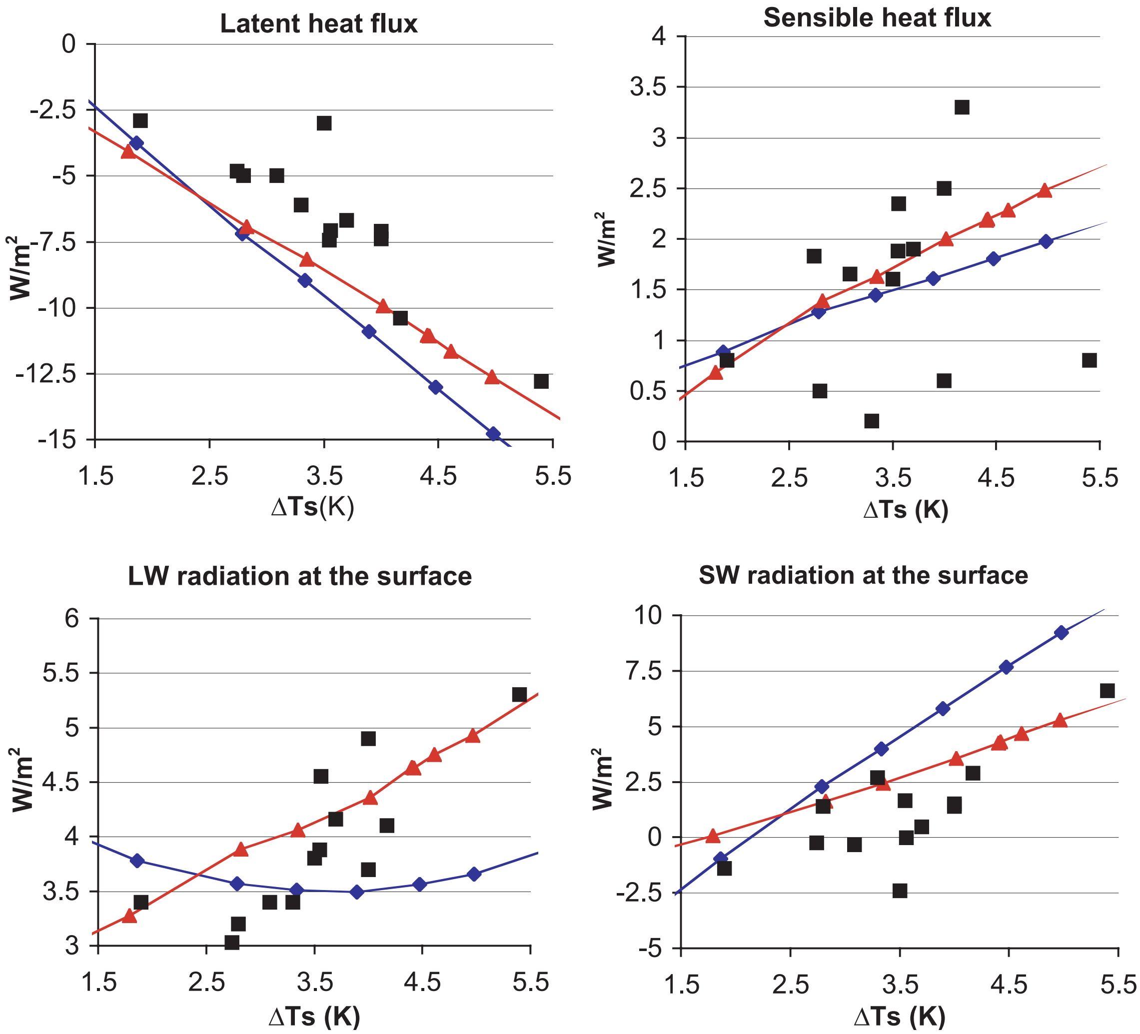

Figure 6. Changes in surface fluxes in equilibrium doubled-CO_2 simulations with different GCMs (squares) and the versions of the MIT climate model with different sensitivities. Diamonds indicate results from the versions with the same k used for all clouds, and triangles from the simulations with k of different signs used for high and low clouds. (See text for details.)

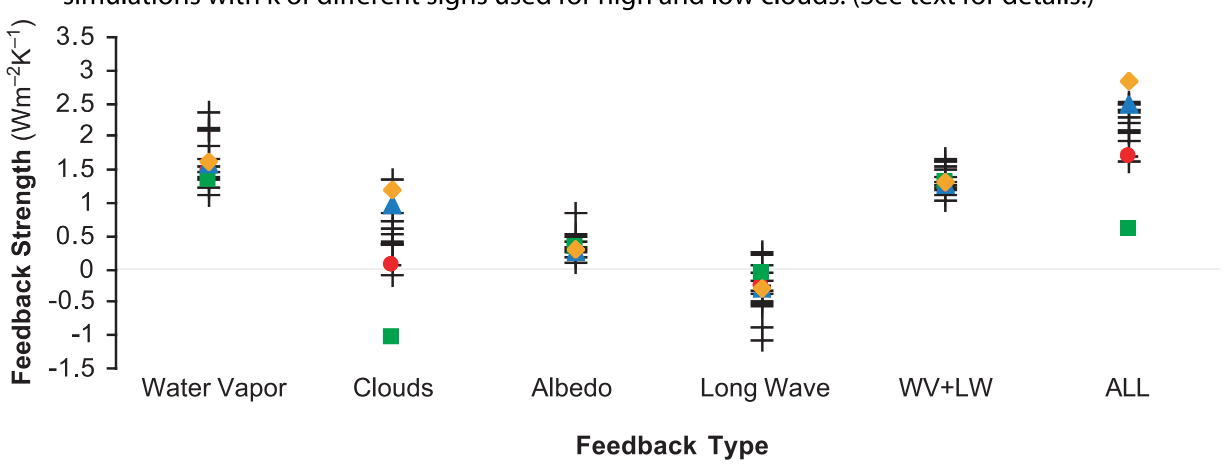

Figure 7. Strengths of feedbacks in different GCMs (+) and in four version of the MIT climate model with different climate sensitivities: 1.39K (squares), 2.18K (circles), 4.5K (triangles), and 7.45K (diamonds). Data for GCMs are from Colman (2003).

In **Figure 7**, the climate feedbacks in four versions of the MIT model are compared with feedbacks in a number of GCMs. Data for different GCMs reported in Colman (2003) were used. Individual feedbacks in all four simulations, with the exception of cloud feedback in the simulation with sensitivity 1.4K, fall in the range shown by GCMs.

5. MODEL SENSITIVITY TO CO_2 INCREASE AS A MEASURE OF MODEL SENSITIVITY TO OTHER FORCINGS

Published results of the simulations with different GCMs do not provide a definitive answer as to whether models' sensitivities to increase in CO_2 concentration (S_{CO2}) reflect sensitivities to other forcings. Models' responses to changes in solar constant and surface albedo are in general consistent with their sensitivities to changes in CO_2 concentration. Comparison of simulations with changes in black carbon or ozone is complicated by differences in simulation design. As such, both Hansen *et al.* (1997) and Cook and Highwood (2004) performed simulations with changes in black carbon; however, magnitudes and vertical structures of those changes were different. Joshi *et al.* (2003) compared responses of three GCMs (UREAD, ECHAM4 and LMD) to an increase in CO_2 concentration, solar constant and upper tropospheric ozone. The magnitudes of changes were chosen such as to produce a forcing of 1 W/m^2 in all cases. For all three cases the strongest response to forcing was produced by the LMD model, and the weakest response by the UREAD model. The differences in sensitivity between models, however, depend on forcing. The ratios of surface warming simulated by the UREAD and LMD models to that simulated by the ECHAM4 for different forcing are shown in **Table 8**. The ratio of sensitivities to ozone is smaller than the ratio of sensitivities to CO_2 for the UREAD model, while larger for the LMD model. Overall, however, ratios for a given model differ by less than 20%.

To see how well sensitivities of the different versions of the MIT model to the CO_2 doubling reflect their sensitivities to other forcings, five additional simulations with sensitivities given in the first row of **Table 9** have been carried out for each forcing. Table 9 shows ratios of the surface air temperature (SAT) changes in those simulations to the SAT change in the simulation with a standard sensitivity for each forcing.

Sensitivity to CO_2 forcing serves as a good measure for sensitivities to the 2%S0 and SRF forcings but noticeably overestimates sensitivities to forcings causing a decrease in surface temperature (–2%S0, 0.5xCO_2, ALB and STRAER; see Table 1). As such, the ratio of the SAT changes in the STRAER simulation with S_{CO2} = 7.45K is about half as large as in the corresponding 2xCO_2 simulation. High sensitivities to changes in CO_2 concentration are primarily caused by large positive shortwave cloud feedback. A significant increase in sea ice

Table 8. Ratios of surface air temperature changes in the simulations with UREAD and LMD GCMs to those in the simulations with ECHAM4.

Forcing	UREAD/ECH	LMD/ECH
CO_2	0.47	1.38
S0	0.38	1.30
O_3	0.41	1.62

Table 9. Ratios of SAT changes in the simulations with low and high sensitivities to those in the simulations with standard sensitivity.

S_{CO2}	0.48	1.39	4.50	5.62	7.45
2xCO2	0.22	0.63	2.06	2.58	3.42
SRF	0.22	0.61	1.93	2.60	3.29
2%S0	0.21	0.6	1.96	2.39	3.39
–2%S0	0.20	0.59	1.53	1.69	1.83
0.5xCO2	0.19	0.59	1.41	1.69	1.82
ALB	0.20	0.59	1.44	1.56	1.82
STRAER	0.16	0.51	1.41	1.52	1.65
10BC	0.19	0.61	1.81	2.15	2.40
LW_BC	0.22	0.61	1.80	2.13	2.47

and snow cover in the last four simulations decreases the effect of changes in clouds on shortwave radiation and therefore decreases the efficiency on an additional cloud feedback. As a result the range of sensitivities to such forcing is narrower than the range of sensitivities to changes in CO_2 concentration.

Since differences in sensitivities between different versions of the MIT climate model are entirely due to differences in cloud feedback, the MIT model will exaggerate difference in sensitivities to positive and negative forcing.[1] At the same time, differences in the strength of cloud feedbacks also account for large part of differences in climate sensitivities between different GCMs (Cess *et al.*, 1990; Colman, 2003), and the interaction between cloud and surface albedo (discussed above) might be relevant for other models.

As shown in Section 3, changes in the radiation fluxes associated with changes in different climate variables, and therefore strengths of different feedbacks in "BC-like" simulations, only partially relates to the surface warming and partially to the warming at the height of the "absorbing" layer. The component of feedbacks not related to surface warming are rather close in magnitude in the simulations with different S_{CO2} (**Table 10**) making the range of the model's sensitivity to "BC-like" forcings smaller than for CO_2 forcing. Model sensitivities calculated from Equation 1 using "corrected" forcing are again close to the sensitivities in corresponding simulations with CO_2 or direct surface forcings.

Changes in precipitation not related to the surface warming (see Section 3) also show very weak dependence on S_{CO2}, as a result the total changes in precipitation depend linearly on the increase in surface temperature (**Figure 8**).

Table 10. Ratios of "corrected" forcings in the "BC-like" simulations with low and high sensitivities to those in the simulations with standard sensitivity.

S_{CO2}	0.48	1.39	4.50	5.62	7.45
LWBC	1.01	0.98	0.89	0.82	0.75
LWCBL3	0.8	1	0.87	0.79	0.75
LWBCL6	0.77	0.77	0.92	0.89	0.79

[1] As a result the MIT model might underestimate impacts of decrease in solar constant or increase in stratospheric aerosol due to volcanic eruptions. It should be kept in mind that forcings used in the above described simulations (see Table 1) are much stronger than the observed ones. For the weaker forcings, this effect will much weaker.

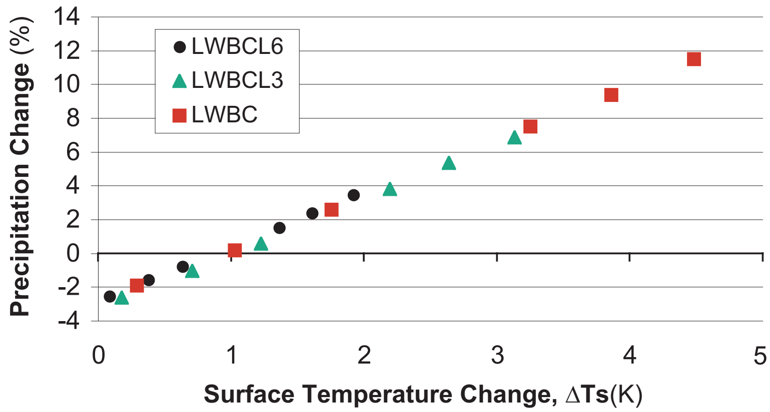

Figure 8. Changes in precipitation in "BC-like" simulations with different sensitivities.

6. CONCLUSIONS

Simulations with the MIT climate model show, similar to the findings of previous studies (*e.g.*, Hansen *et al*., 1997; Cook and Highwood, 2004), a strong dependence of the model response on vertical structure of the imposed forcing. Heating in the lowest 1500 meters produces much stronger surface warming than an equivalent heating of the upper layers. Such dependency of surface warming on the altitude of heating is explained by differences in cloud and joint water vapor/lapse rate feedbacks. If, however, changes in radiation fluxes associated with the above surface warming are treated as a "semi-direct" forcing, then the total forcing provides a good measure for the increase in surface temperature.

Simulations with versions of the MIT model with different strengths of cloud feedback show that model sensitivity to the increase in CO_2 concentration reasonably well characterizes the model's sensitivity to other positive forcing with similar vertical structure. In the case of the forcings leading to surface cooling, an increase in the strength of cloud feedback is less efficient due to an increase in sea ice extent and snow cover, and associated with that, an increase in surface albedo. Since differences in cloud feedback are one of the main reasons for the differences in sensitivities between different GCMs, this implies that the range of the models' responses to such forcing as increase in stratospheric aerosol or decrease in solar constant might be narrower that the range of responses to CO_2 increase.

Sensitivity to changes in the CO_2 concentration is defined by the strength of climate feedbacks related to surface warming. A distinguishing feature of the simulations with "black-carbon-like" forcings is a presence of additional feedbacks related to the warming at the location of "absorbing" layer ("semi-direct" forcing). Therefore, sensitivities defined through doubled-CO_2 simulations may not provide good estimates for the sensitivities to forcing with different vertical structures. As such, the range of the MIT model responses to changes in black carbon concentration and "BC-like" forcings is also smaller than to changes in CO_2. The latter is explained by "semi-direct" forcings having similar magnitude in the simulations with different strengths of cloud feedback. Large differences, however, occur for values of S_{CO2} outside of the range produced by existing GCMs.

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