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A Comparison of the Behavior of AOGCMs in Transient Climate Change Experiments

Andrei P. Sokolov, Chris E. Forest and Peter H. Stone

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A Comparison of the Behavior of Different AOGCMs in Transient Climate Change Experiments

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Abstract

The transient response of both surface air temperature and deep ocean temperature to an increasing external forcing strongly depends on climate sensitivity and the rate of the heat mixing into the deep ocean, estimates for both of which have large uncertainty. In this paper we describe a method for estimating rates of oceanic heat uptake for coupled atmosphere/ocean general circulation models from results of transient climate change simulations. For models considered in this study, the estimates vary more than threefold. Nevertheless, values for all models fall in the 5–95% interval of the range implied by the climate record for the last century.

The MIT 2D climate model, with an appropriate choice of parameters, matches changes in surface air temperature and sea level rise simulated by different models. It also reproduces the overall range of changes in precipitation.

Contents

1. INTRODUCTION

At the present time, coupled atmosphere-ocean general circulation models (AOGCMs) are widely used for making projections of possible future climate change. However, results produced by different AOGCMs differ significantly even for similar changes in external forcing. For example, in simulations with 1% per year increase in CO_2 concentration, performed in the framework of the Coupled Models Intercomparison Project (http://www-pcmdi.llnl.gov/cmip/index.html), the increase in surface air temperature (SAT) at the time of CO_2 doubling (an average for years 61–80) simulated by different models ranges from 1.32 ^{o}C to 2.15 ^{o}C (Covey *et al*., 2000).

The transient response produced by a given model is, to a large part, determined by two characteristics of the model: sensitivity to an external forcing and the rate of heat uptake by the ocean. While sensitivities for many AOGCMs are known and given in the literature, differences in the rates of oceanic heat uptake are not well estimated. The ratio of the SAT increase at the time of CO_2 doubling to the equilibrium model sensitivity, which is often used to compare transient responses of different AOGCMs (see for example, Murphy and Mitchell, 1995), depends on both the rate of oceanic heat uptake and model sensitivity. In upwelling-diffusion models, a number of parameters, such as a mixed layer depth, the upwelling rate, a diffusion coefficient and so on, are varied to fit the behavior of different AOGCMs (Wigley and Raper, 1993; Cubasch *et al*., 2001). The use of multiple parameters makes it difficult to compare the rates of heat uptake by different models. In this study, we obtain quantitative estimates for the oceanic heat uptake by choosing parameters of the MIT 2D climate model to match behavior of different AOGCMs. Then, the effective heat diffusivity of the MIT model provides a measure of the rate of the heat uptake by the deep ocean for those models. This study was conducted as a part of subproject #20 of CMIP2.

2. MODEL DESCRIPTION

The atmospheric component of the MIT 2D climate model (Sokolov and Stone, 1998) is a zonal averaged statistical-dynamical model developed from the GISS AGCM (Hansen *et al*., 1883). It includes parameterizations of all the main physical processes in the atmosphere and therefore, can reproduce major feedbacks. It also includes parameterizations for atmospheric heat, moisture, and momentum transports by large-scale eddies (Stone and Yao, 1987, 1990). For any given AOGCM, model sensitivity, as well as the rate of the oceanic heat uptake, depend on how different feedbacks are depicted by the model, which, in turn, is defined by a large number of factors, such as parameterizations of different physical processes, horizontal and vertical resolutions, and so on. In contrast, the sensitivity (S) of the MIT 2D model can be specified by changing the strength of the cloud feedback. Namely, the amount of clouds used in radiative transfer calculations is defined as $C = C_o(1+k\Delta T_s)$, where C_o is the simulated cloud cover and ΔT_s is the deviation of global mean SAT from its value in an equilibrium present-day climate simulation (Hansen *et al*., 1993). It was shown by Sokolov and Stone (1998) that the dependence of changes in different climate variables, such as precipitation, surface fluxes and so on, on climate sensitivity shown by the MIT model is similar to the dependence found in equilibrium climate change simulations with different AGCMs.

The ocean component of the MIT 2D climate model consists of a Q-flux mixed layer model with a deep ocean diffusive model beneath it. The mixed layer depth is prescribed from observations as a function of season and latitude. In addition to the temperature of the mixed layer, the model also calculates the averaged temperature of the seasonal thermocline and the temperature at the annual maximum depth of the mixed layer (Russell *et al*., 1985). In contrast with conventional diffusive models, diffusion in the MIT model is not applied to temperature itself but to the temperature difference from its values in a present-day climate simulation (Hansen *et al*., 1984; Sokolov and Stone, 1998). In our model, diffusion represents a cumulative effect of the mixing of heat by all physical processes and therefore, the values of the diffusion coefficients are significantly larger than those used in sub-grid scale diffusion parameterizations in OGCMs. The values of effective diffusion coefficients calculated from data on tritium mixing into deep ocean (Hansen *et al*., 1984) vary from 0.2 cm^2/s in tropics to about 10 cm^2/s in high latitudes with a global averaged value of 2.5 cm^2/s. The rate of heat penetration into the deep ocean is varied by multiplying diffusion coefficients by the same factor at each latitude thereby preserving the spatial structure of the heat uptake. Despite the ocean component's simplicity, the MIT model can reproduce the evolution of different AOGCMs in typical climate change scenarios for about 100–150 years, in terms of global mean SAT and the sea level rise due to thermal expansion of the deep ocean (Sokolov and Stone, 1998).

3. ESTIMATING RATES OF HEAT UPTAKE FOR DIFFERENT AOGCMS

A number of climate change simulations with different coupled AOGCMs have been carried out in the second stage of the Coupled Model Intercomparison Project (CMIP2) (http://www-pcmdi.llnl.gov/cmip/index.html). In these simulations, models were forced by 1% per year increase in the atmospheric CO_2 concentration for 80 years. To compare behavior of different AOGCMs, we obtain versions of the MIT 2D climate model that fit the response of the models in question. The global averaged values of diffusion coefficients (K_v) used in the fits for different AOGCMs give a measure for their rate of oceanic heat uptake.

Apart from the region of low climate sensitivity ($S < 1$ °C), SAT change and sea level rise due to thermal expansion of the ocean are unequivocally defined by S and K_v (**Figure 1**). Thus, a fit for a given AOGCM can be estimated based on the data on surface warming and thermal expansion of the ocean. However, if the value of the model's sensitivity is already known, then the value of K_v can be chosen so that the transient change of SAT for this model is reproduced by the MIT 2D model with the same sensitivity. Data on sea level rise then can be used to check the quality of the fit. We used the latter approach whenever possible.

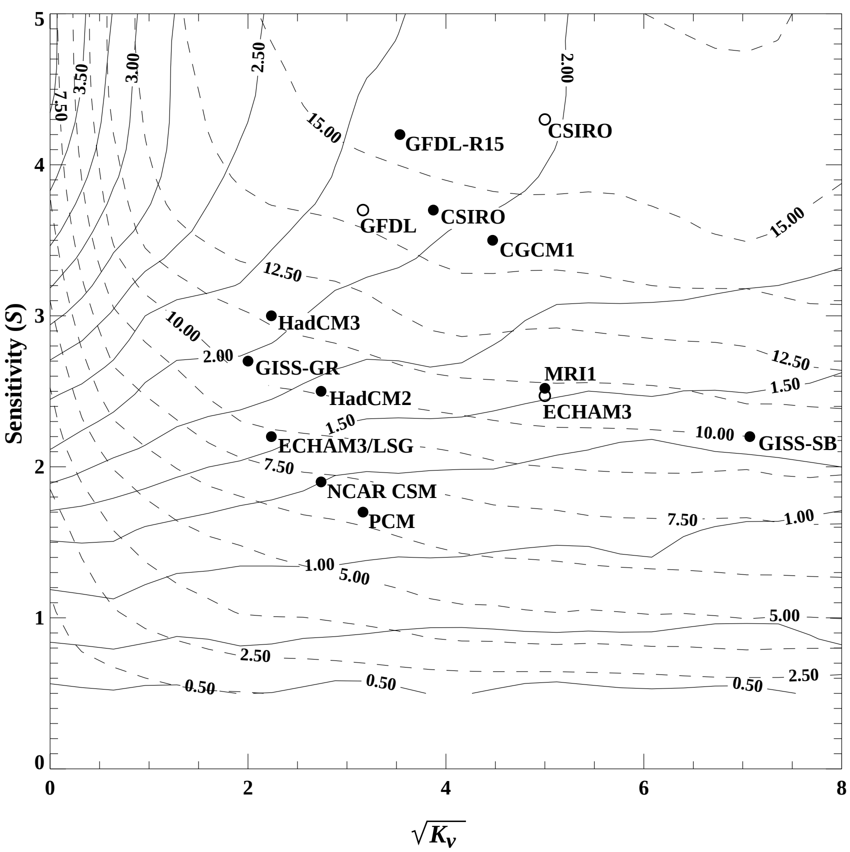

Figure 1. Changes in surface air temperature and sea level rise due to thermal expansion of the ocean at the time of CO_2 doubling. See text for details.

Sensitivity for a given AOGCM is usually defined as the equilibrium surface warming (ΔT_{eq}) simulated by the corresponding atmospheric model coupled to a mixed layer ocean model in response to the doubling of atmospheric CO_2 concentration. It varies from about 2 ^oC to about 5 ^oC among existing AOGCMs (Cubasch *et al*., 2001). The estimates for an equilibrium sensitivity from simulations with coupled AOGCMs are available to date only for the HadCM2 (Senior and Mitchel, 2000) and the GFDL_R15 (Stouffer and Manabe, 1999) models. In both cases they are somewhat different from those obtained in the simulations with mixed layer ocean models.

It was noticed by Murphy (1995) that the sensitivity of a coupled AOGCM changes with time due to changes in the strength of different atmospheric feedbacks.[1] The energy balance of the climate system can be described by the following simple equation:

$$C \frac{\partial \Delta T(t)}{\partial t} = F(t) - \lambda \Delta T(t) , \quad (1)$$

where C is the heat capacity of the system, $F(t)$ is an external forcing, ΔT is the change in surface temperature and λ is a feedback parameter. In equilibrium, $\lambda_{eq} = F_{2xCO2}/\Delta T_{eq}$, where F_{2xCO2} is a forcing due to CO_2 doubling. In a transient run, a time-dependent effective feedback parameter can be estimated as follows:

$$\lambda_{eff}(t) = \frac{F(t) - R_{toa}(t)}{\Delta T(t)} , \quad (2)$$

where $R_{toa}(t)$ is the net radiative flux at the top of the atmosphere. An effective climate sensitivity, $\Delta T_{eff,}$ is then defined as what the equilibrium surface warming due to CO_2 doubling would be if $\lambda = \lambda_{eff}$, $\Delta T_{eff} = F_{2xCO2}/\lambda_{eff}$. The values of the effective sensitivity at the time of CO_2 doubling for some AOGCMs used in the CMIP2 simulations are given in Cubasch *et al*. (2001) and are shown in **Table 1**. As can be seen, the effective sensitivity at the time of CO_2 doubling is usually smaller than ΔT_{eq}, and for some models significantly smaller.

Table 1. Values of equilibrium and effective climate sensitivities at the time of CO_2 doubling from Cubasch *et al.* (2001). Values of ΔT_{eq} are from simulations with mixed-layer ocean models, while ΔT_{eff} are from transient simulations with coupled AOGCMs.

Model	ΔT_{eq}	ΔT_{eff} **at 2x**CO_2
CGCM1	3.5	3.6
CSIRO	4.3	3.7
ECHAM3/LSG	2.5	2.2
GFDL_R15	3.7	4.2
HadCM2	4.1	2.5
HadCM3	3.3	3.0
MRI1	4.8	2.6
NCAR_CSM	2.1	1.9

[1] Changes in the model sensitivity described by Senior and Mitchell (2000) occurring after a few hundreds years of integration and associated with changes in the deep ocean circulation are not relevant when results of relatively short-term simulations are analyzed.

Satisfactory fits have been obtained for a number of the AOGCMs using equilibrium climate sensitivities (Sokolov and Stone, 1998). However, for some models used in CMIP2 simulations, thermal expansion was overestimated by the versions of the MIT model with *S* equal to the model's equilibrium climate sensitivity, even so, they fit the SAT changes. For example, very large effective diffusion coefficients (K_v = 500 cm^2/s) are required to reproduce changes in SAT simulated by the MRI1 AOGCM (**Figure 2a**) when the model's equilibrium climate sensitivity of 4.8 ^{o}C is used. However, the MIT climate model with these parameters produces a significantly larger sea level rise (Fig. 2b).[2] At the same time, the MIT model with *S* = 2.6 ^{o}C and K_v = 50 cm^2/s reproduces changes in both SAT and sea level.

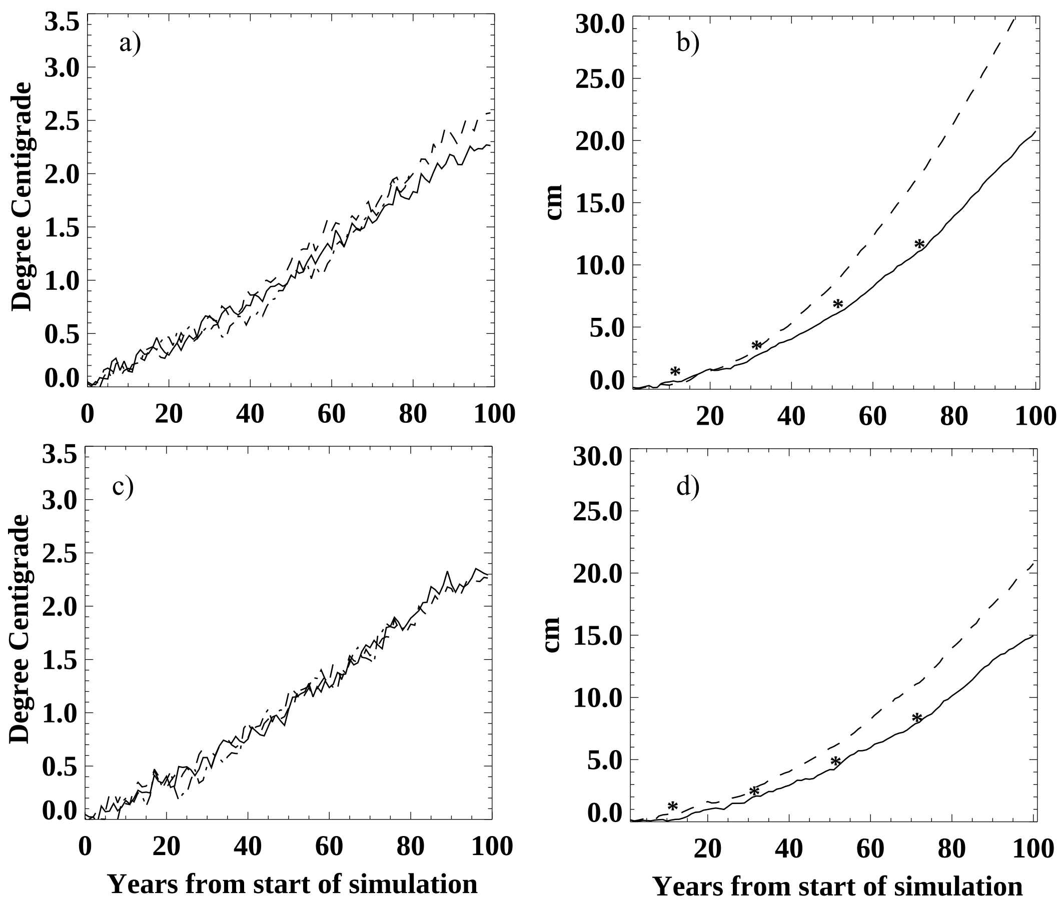

Figure 2. Changes of annual mean global mean surface air temperature and sea level (thermal expansion) in simulations with the MRI1 (a,b) and ECHAM3/LSG (c,d) AOGCMs and in simulations with the versions of the MIT 2D Climate Model with effective (solid lines) and equilibrium (dashed lines) climate sensitivities. Data from CMIP2 simulations with AOGCMs are shown by dashed-dotted line (SAT) and by * (sea level).

[2] Unfortunately, while changes in SAT from these simulations are available on an annual basis, sea level rise due to thermal expansion of the ocean is not. The data required to calculate thermal expansion were saved as a 20 year mean for four consecutive segments of the simulations. In this study we used data on sea level rise for these four periods provided by Sarah Raper (Raper *et al*., 2001).

Using the effective sensitivity, instead of an equilibrium one, leads to significantly better simulation of the oceanic thermal expansion not only for the MRI1 but also for the ECHAM3/LSG AOGCM in spite of the small difference between the two sensitivities for the latter model. On the other hand, fits with effective and equilibrium sensitivities give very close results for the CSIRO and GFDL_R15 models (**Figure 3**). Positions of the final fits for different AOGCMs using $S = \Delta T_{eff}$ are shown in Figure 1 by filled circles. Positions of the versions of the MIT model which reproduce changes in SAT for ECHAM3/LSG, CSIRO and GFDL_R15 using their equilibrium sensitivities are shown by open circles. Due to the weak dependence of changes in SAT on K_v for low climate sensitivities, the two fits for the ECHAM3/LSG model have significantly different rates of oceanic uptake. Sea level rise, on the contrary, is rather sensitive to changes in K_v in this region of parameter space. This, together with the relatively small increase in sea level projected by the ECHAM3/LSG model explains the noticeable difference between this model's fits with equilibrium and effective sensitivities. The opposite is true for both the CSIRO and the GFDL_R15 models. It should be noted that the difference in sea level rise projections by fits with different sensitivities increases with time. In general, the use of an effective sensitivity instead of an equilibrium one leads to better simulation of sea level rise.

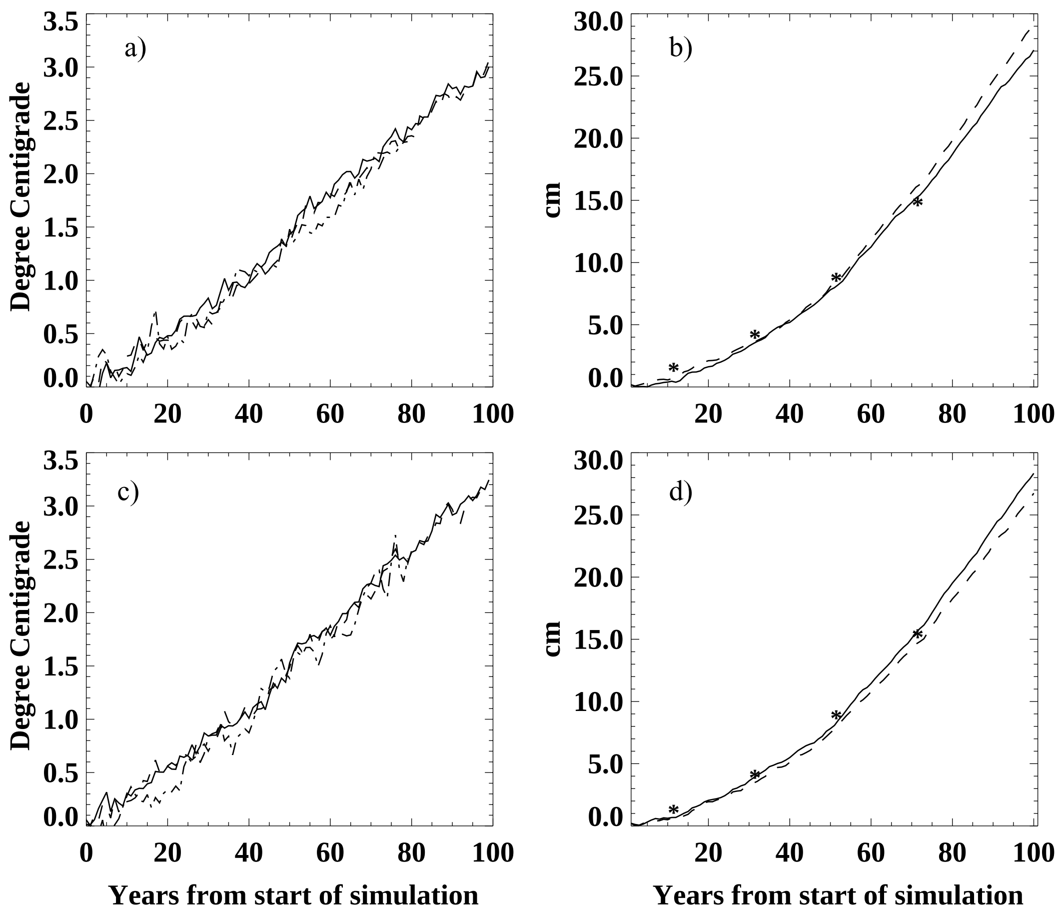

Figure 3. The same as Figure 2 but for the CRISCO Mk2 (a,b) and GFDL_R15 (c,d) AOGCMs.

Table 2. Adjusted radiative forcing due to CO_2 doubling for different models.

Model	**Forcing (W/m²)**
CSIRO	3.45
HadCM2	3.47
HadCM3	3.74
NCAR_CSM	3.60
PCM	3.60
GISS and MIT 2D	3.84

In all simulations discussed above, the MIT climate model was forced by radiative forcing calculated by its radiation scheme (in contrast with energy balance models where the radiative forcing is prescribed). It has been shown, however, that different models produce different forcings for the same increase in the CO_2 concentration (Cess *et al.*, 1993). The values of the adjusted radiative forcing[3] due to CO_2 doubling for some models are given in **Table 2**.

A number of additional simulations have been carried out to evaluate the impact of these differences. The differences in forcing were taken into account in the following way. As is well known, radiative forcing increases linearly with an exponential increase in CO_2, namely $F(t) = \kappa\alpha t$, where α is a rate of CO_2 increase and κ is a coefficient different for different models. A value of κ for a given model is defined by the details of its radiation code (for the MIT 2D model $\kappa = 5.35$) and cannot be changed. Therefore, we changed α such that the forcing averaged over years 61–80 matched a given model's value. However, if differences in forcing are taken into account, the 2D model's sensitivity (S) must also be changed to match the "specific" sensitivity, that is an equilibrium SAT increase due to forcing of 1 W/m^2, of a given. The values of S used in simulations with 1% per year increase in CO_2 (Table 1, 2nd column) are defined as a surface warming in response to CO_2 doubling or, more exactly, to the forcing produced by CO_2 doubling in the MIT model (that is, 3.84 W/m^2). For example, S = 3.7 ^{o}C, matching the equilibrium sensitivity of the CSIRO AOGCM, corresponds to a warming of 0.96 ^{o}C/(W/m^2) for the MIT 2D model while "specific" sensitivity of the CSIRO AOGCM is 1.07 ^{o}C/(W/m^2). Therefore, a climate sensitivity of 4.12 ^{o}C should be used in the simulation with the MIT 2D model to match a "specific" sensitivity of the CSIRO AOGCM.

The CSIRO and HadCM2 AOGCMs produce forcing most different from that of the 2D model (Table 2). However, simulations with corrected values of forcings and sensitivities even for these models (**Figure 4**) show small differences compared to the simulations with the original sensitivities and forcing. Such a small impact of different forcing on the results of simulations with increasing CO_2 can be explained through simple analysis of equation (1). For linear forcing, equation (1) has an analytical solution under the assumption that C is fixed. Namely:

$$\Delta T_s(t) = \gamma S(t - \tau(1 - e^{-\frac{t}{\tau}})), \qquad (3)$$

where $\gamma = \kappa\alpha$, $S = \lambda^{-1}$ and $\tau = SC$.

[3]Adjusted refers to the radiative imbalance at the tropopause after the stratospheric temperatures have adjusted to the new CO_2 concentration. This adjusted forcing must be used in energy balance models (EBMs) to reproduce the behavior of AOGCMs (Cubasch *et al.*, 2001; Raper *et al.*, 2001).

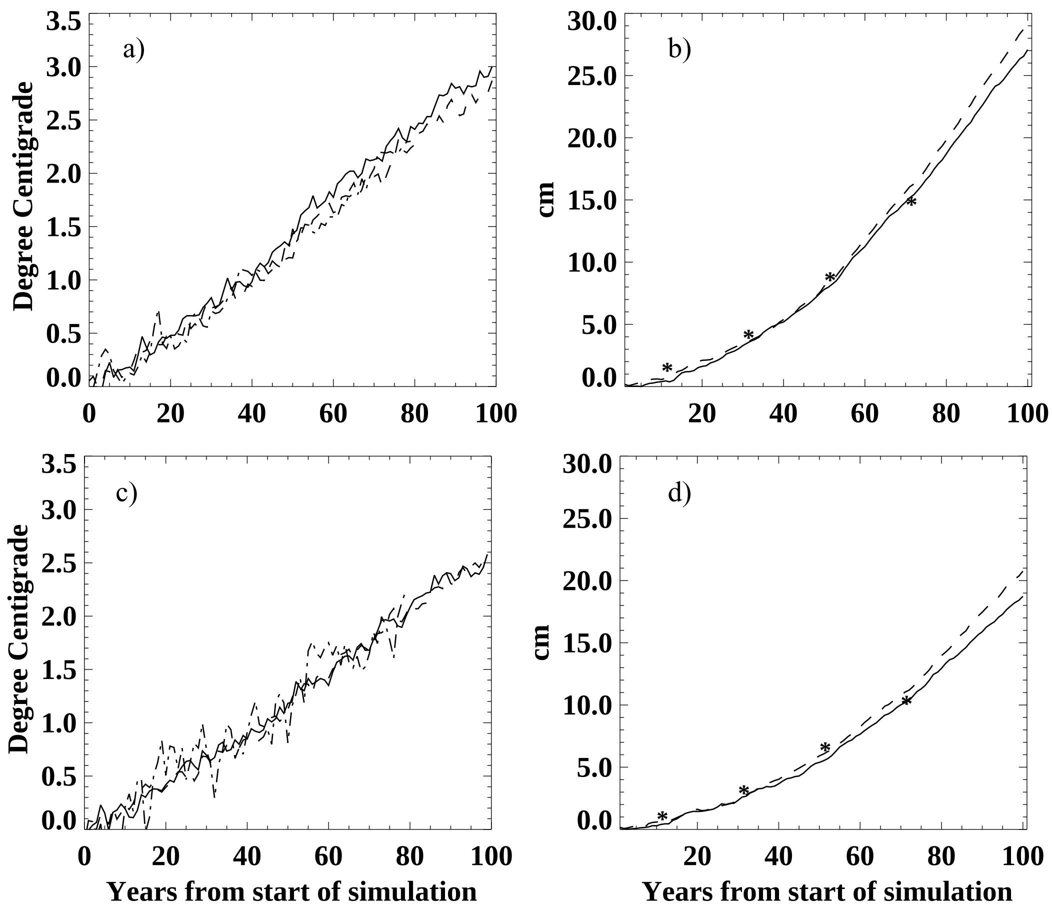

Figure 4. Changes of annual mean global mean surface air temperature and sea level (thermal expansion) in simulations with the CSIRO (a,b) and HadCM2 (c,d) AOGCMs and in simulations with the versions of the MIT 2D Climate Model with corrected (dashed lines) and uncorrected (solid lines) forcing. Data from CMIP2 simulations with AOGCMs are shown by dashed-dotted line (SAT) and by * (sea level). Data from CMIP2 simulation with AOGCMs are shown by dashed-dotted line (SAT) and by * (sea level).

However, for equation (1) to be a correct equation for the change in surface air temperature, C should be the heat capacity of the part of the deep ocean affected by warming at a time t but not the heat capacity of the whole ocean. The former is proportional to the depth of heat anomaly penetration, which for a diffusive model is proportional to $\sqrt{K_v * t}$ (Hansen *et al.*, 1985). While equation (3) is not an exact solution of equation (1) for time dependent C, it approximates a numerical solution of equation (1) rather well with τ proportional to $S\ \sqrt{K_v * t}$. While values of γ and S are different in simulations with corrected and uncorrected forcings, their product is the same in both cases. As a result, the difference in ΔT_s is relatively small in spite of difference in τ. As could be expected, the difference is large for the CSIRO AOGCM due to a larger τ. Analogous simulations with other models have shown that taking into account differences in forcing between different AOGCMs does not noticeably affect estimates of the rates of oceanic heat uptake. Because data on radiative forcing are not available for all models, the estimates from simulations with 1% per year increase in CO_2 concentration were used. Fits for the

GISS_GR (Russell *et al*., 1995) and the GISS_SB (Sun and Bleck, 2001) AOGCMs were obtained based on the data on SAT and thermal expansion, provided by the models' authors, without prior knowledge of the models' sensitivities. Fits for some models used in CMIP2 simulations were not obtained due to absence of data on sea level rise.

As follows from above, the most natural measure for the rate of oceanic heat uptake for the MIT 2D model is $\sqrt{K_v}$. For models given in **Table 3**, $\sqrt{K_v}$ varies from 2.0 to 7.1 cm/s$^{1/2}$. In **Figure 5** a probability density function (PDF) for $\sqrt{K_v}$ calculated from data for models is compared with the one based on comparison with observations (Forest *et al*., 2001). The PDF for models was obtained by fitting β distributions to data from Table 3. Data for all models were weighted equally. Though shapes of the two PDFs are different, values of K_v for all models fall into the 5–95% interval suggested by observations. The means/medians of the two distributions are also not very different, 3.49/3.33 cm/s$^{1/2}$ and 4.20/4.40 cm/s$^{1/2}$ for models and observations, respectively. It should be noted that, because the observations do not place an upper bound on K_v, a subjective bound of K_v = 64 cm^2/s was imposed. For a different choice of an upper bound the values of fractals would be somewhat deferent. A PDF for the rate of oceanic heat uptake, which is a key factor for projecting future climate change, based on estimates derived from comparison with both observations and different AOGCMs was used by Webster *et al.* (2001).

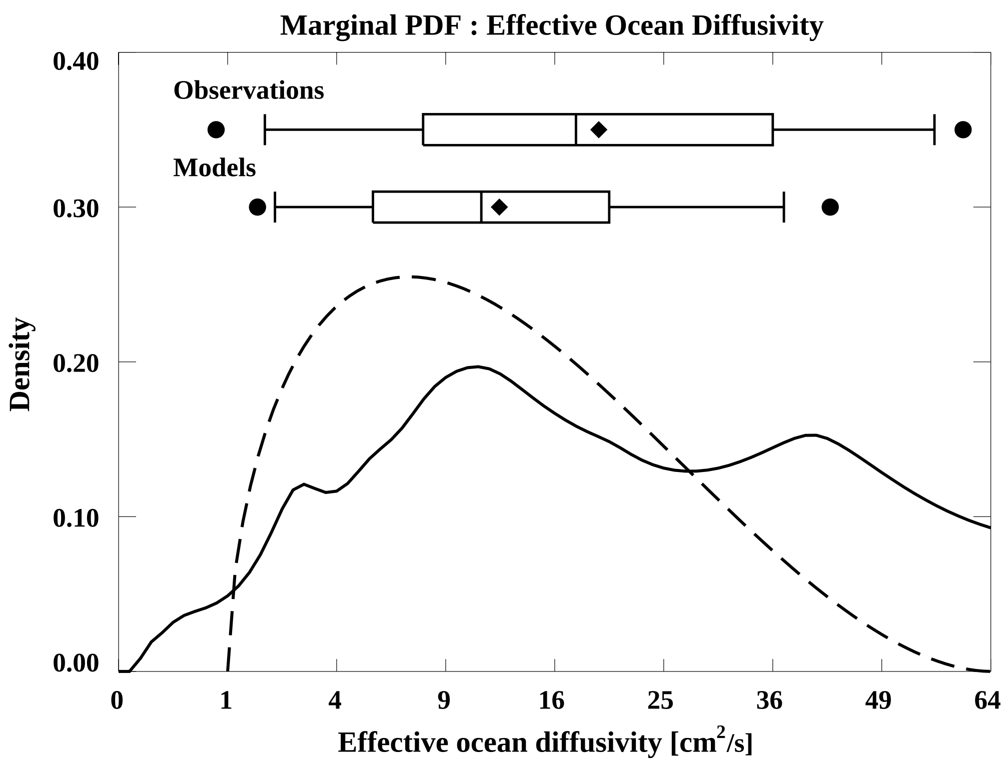

Figure 5. Probability density functions for the rate of oceanic heat uptake from models (dashed) and observations (solid). The whisker plots show the 2.5–97.5% (dots), 5–95% (vertical bars on ends), and 25–75% (box) probability ranges along with the median (bar within box) and mean (diamond) for each distribution.

Table 3. Parameters of the versions of the MIT climate model simulating behavior of different AOGCMs. The values of τ are at the time of CO_2 doubling.

	Parameters of corresponding versions of the 2D model			
Models	S (°C)	K_v (cm²/s)	$\sqrt{K_v}$ (cm/s$^{1/2}$)	τ (years)
CGCM1	3.6	20	4.47	133.8
CSIRO	3.7	15	3.87	119.6
ECHAM3/LSG	2.2	5	2.24	57.7
GFDL_R15	4.2	12.5	3.54	123.8
GISS_GR	2.7	4.0	2.0	45.2
GISS_SH	2.2	50.0	7.07	130.5
HadCM2	2.5	7.5	2.74	56.9
HadCM3	3.0	5.0	2.24	56.1
MRI1	2.6	25.0	5.0	108.8
NCAR_CSM	1.9	7.5	2.74	39.3
PCM	1.7	10.0	3.16	50.2

4. CHANGE IN PRECIPITATION

As shown above, the MIT 2D climate model with an appropriate choice of climate sensitivity and an effective diffusion coefficient can reproduce changes in SAT and sea level projected by different AOGCMs. Because the MIT climate model is used as a component of the MIT Integrated Global System Model (IGSM) (Prinn *et al.*, 1999), it is also important to know how it simulates transient changes in other climate variables. Precipitation is of particular interest because it is used as an input by both the Terrestrial Ecosystem Model (Xiao *et al.*, 1997) and the Natural Emission Model (Liu, 1997), which are included in the IGSM

Changes in precipitation, even on a global scale, are not unequivocally defined by global characteristics of a given model, but depend on details of the physical parameterizations. Thereby, a version of the MIT model matching transient changes in SAT and sea level simulated by a particular AOGCM does not necessarily reproduce changes in precipitation for the same model. For example, the MIT model simulates rather well changes in precipitation for the CSIRO and ECHAM3/LSG AOGCMs (**Figure 6a**), but significantly overestimates the increase in precipitation for the CGCM1 model and underestimates it for the MRI1 model (Figure 6b). **Figures 7** reveals a strong positive correlation between changes in precipitation and SAT in different simulations with the MIT climate model. In contrast, a noticeably weaker correlation exists between changes in those two variables as simulated by different AOGCMs. While the results of simulations with the MIT model almost fall on a straight line, the results from AOGCMs are more scattered. Covey *et al*. (2000) showed that the correlation is also weak when results of all CMIP2 simulations are compared. While precipitation increases (in terms of the global average) with an increase in SAT in all simulations, the rate of such an increase for a given model is mainly defined by parameterizations of different physical processes, such as convection, cloud formation, or calculation of surface fluxes (Washington and Meehl, 1993). Because the only difference between different versions of the MIT model is the strength of cloud feedback, the above-mentioned strong correlation between changes in SAT and precipitation for the MIT model simulations is not surprising. As a result, the MIT climate model cannot simulate some particular climate change regimes, such as cold and wet or hot and dry climates. Nevertheless, it reproduces the range of increases in precipitation produced by AOGCMs.

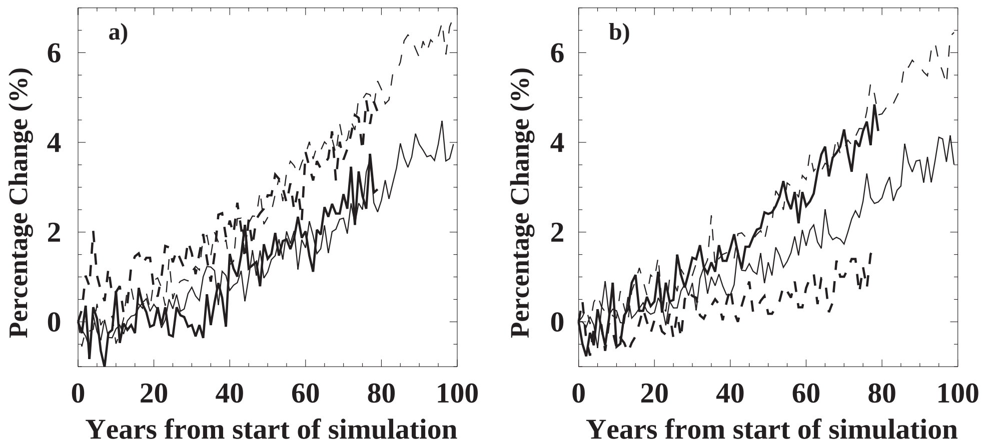

Figure 6. Changes of annual mean global mean precipitation in simulations with AOGCMs (thick lines) and with the matching versions of the MIT 2D Climate Model (thin lines); a) for CSIRO (dashed lines) and ECHAM3 (solid) models, b) for MRI (dashed) and CGCM1 (solid) models.

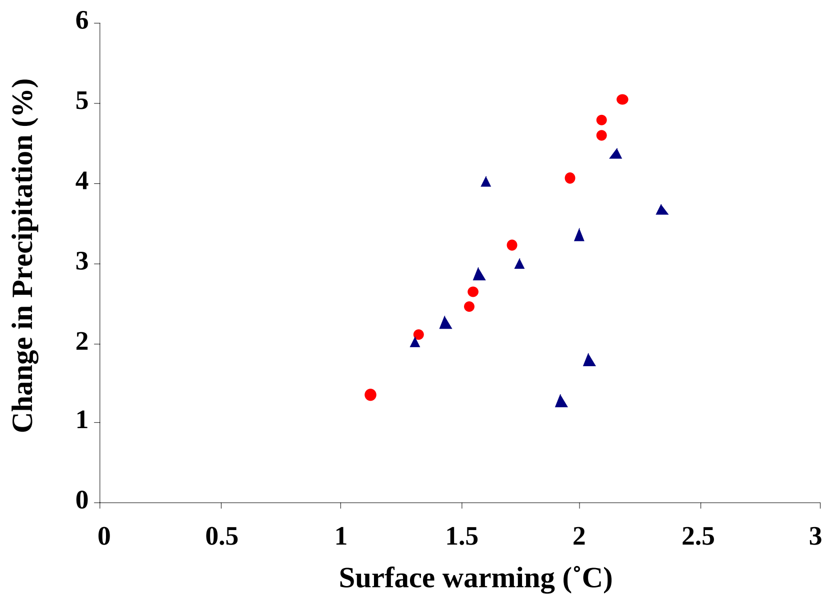

Figure 7. Percentage change in globally and annually averaged precipitation as a function of global mean warming at the time of doubling of CO_2 as produced by different AOGCMs (triangles) from CMIP2 and corresponding versions of the MIT 2D model (circles).

5. CONCLUSIONS

The MIT 2D climate model with an appropriate choice of parameters defining the model's sensitivity and the rate of oceanic heat uptake can successfully reproduce both an increase in surface air temperature and sea level rise due to thermal expansion of the deep ocean projected by a given AOGCM. The rate of heat uptake by the deep ocean in the MIT model is defined by one parameter, namely the global averaged value of an effective diffusion coefficient. This provides quantitative estimates of the strength of oceanic heat uptake for different AOGCMs.

Use of an effective climate sensitivity at the time of CO_2 doubling, instead of an equilibrium sensitivity, leads to better fits and for some models to significantly different estimates of oceanic heat uptake. At the same time, taking into account differences in the radiative forcing between different AOGCMs does not noticeably affect those estimates.

Estimated values of effective diffusion coefficients for AOGCMs considered in this study differ by more than factor of three (in terms of $\sqrt{K_v}$), and this introduces considerable uncertainty in long-term projections of climate change. It should be noted that the values for all models fall within the 5–95% interval of the range derived from comparisons with the 20^{th} century climate record (Forest *et al*., 2001).

Different versions of the MIT climate model show stronger correlation between changes in SAT and global averaged precipitation than simulated by AOGCMs. Nevertheless, the MIT climate model, while not matching results of some models, does capture the range of increases in precipitation produced by AOGCMs.

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