Comparing Oceanic Heat Uptake in AOGCM 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 a method for estimating rates of oceanic heat uptake for coupled atmosphere–ocean general circulation models from results of transient climate change simulations is described. For models considered in this study, the estimates vary by a factor of 2½. Nevertheless, values of oceanic heat uptake for all models fall in the range implied by the climate record for the last century. It is worth noting that the range of the model values is narrower than that consistent with observations and thus does not provide a full measure of the uncertainty in the rate of oceanic heat uptake.

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% yr^{-1} increase in CO_2 concentration, performed in the second stage of the Coupled Models Intercomparison Project (CMIP2; 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° to 2.15°C (Covey et al. 2000).

The transient response produced by a given model to a given forcing is, to a large extent, 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, e.g., 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. However, with the Massachusetts Institute of Technology (MIT) 2D climate model it is possible to match the behavior of different AOGCMs by varying just two parameters: climate sensitivity and effective diffusivity for ocean heat anomalies (Sokolov and Stone 1998). The effective diffusivity of the MIT model provides a measure of the rate of the heat uptake by the deep ocean for AOGCMs. Some of the climate characteristics that have to be specified in energy balance models, such as land–ocean temperature gradient and the rate of atmospheric heat exchange between Northern and Southern Hemispheres, are simulated by the MIT model.

The "ocean heat uptake efficiency"[1] used by Raper et al. (2002) is well correlated with our effective diffusion coefficient. The former, however, depends on time and a particular forcing scenario. Therefore, ocean heat uptake efficiencies for different AOGCMs can be compared only if calculated from simulations with identical forcings. In contrast the effective diffusion coefficients needed in the MIT model to match results of different models are independent of the forcing and can, therefore, be estimated from simulations with different forcings.

It should be noted that in our model, diffusion is used to simulate mixing of heat anomalies by all processes. Therefore, our effective diffusion coefficients cannot be directly compared with coefficients used in either

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[1] Ocean heat uptake efficiency is a ratio of global averaged heat flux into ocean to an increase in surface air temperature (Gregory and Mitchell 1997).

OGCMs or upwelling–diffusion models or variants thereof. This study was conducted as a part of subproject number 20 of CMIP2.

2. Model description

The atmospheric component of the MIT 2D climate model (Sokolov and Stone 1998) is a zonally averaged statistical–dynamical model developed from the Goddard Institute for Space Studies Atmospheric General Circulation Model (GISS AGCM; Hansen et al. 1983). 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, depends 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 of the MIT 2D model (S) can be specified by changing the strength of the cloud feedback. Namely, the amount of clouds used in radiative transfer calculations is defined as $C_* = C_0 (1 + k\Delta T_S)$, where C_0 is the cloud cover simulated by the model 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 on climate sensitivity of changes in the global mean values of different climate variables, such as precipitation, and surface fluxes as simulated by the MIT model is similar to the dependence found in equilibrium climate change simulations with four versions of the U.K. Met Office (UKMO) AGCM with different cloud parameterizations (Senior and Mitchell 1993). The model also reasonably well simulates distributions of zonal mean changes in, for example, surface air temperature, as well as the vertical structure of warming.

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 upwelling–diffusion models, diffusion in our deep ocean 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). Because diffusion represents a cumulative effect of the mixing of heat by all physical processes, the values of the diffusion coefficients are significantly larger than those used in subgrid-scale diffusion parameterizations in OGCMs or in upwelling–diffusion models. Although

the effective diffusion coefficients based on diapycnal and isopycnal mixing alone are different for temperature and nontemperature tracers, being much larger for the latter (Harvey 2001), the contribution to the effective K_v from convective mixing should be the same for all tracers and is dominant at high latitudes (Harvey 1995). Effective diffusion coefficients used in the MIT model are based on coefficients calculated from data on tritium mixing into the deep ocean (Hansen et al. 1984), which vary from 0.2 cm^2 s^{-1} in the Tropics to about 10 cm^2 s^{-1} in high latitudes with a global average value of 2.5 cm^2 s^{-1}. 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 framework of CMIP2. In these simulations, models were forced by 1% yr^{-1} 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 (Fig. 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. Sea level rise data then can be used to check the quality of the fit. We used the latter approach whenever possible.

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° to about 5°C 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 Hadley Centre second generation coupled ocean–atmosphere GCM (HadCM2; Senior and Mitchell 2000) and the Geophysical Fluid Dynamics Laboratory (GFDL) R15 (Stouffer and Manabe 1999)
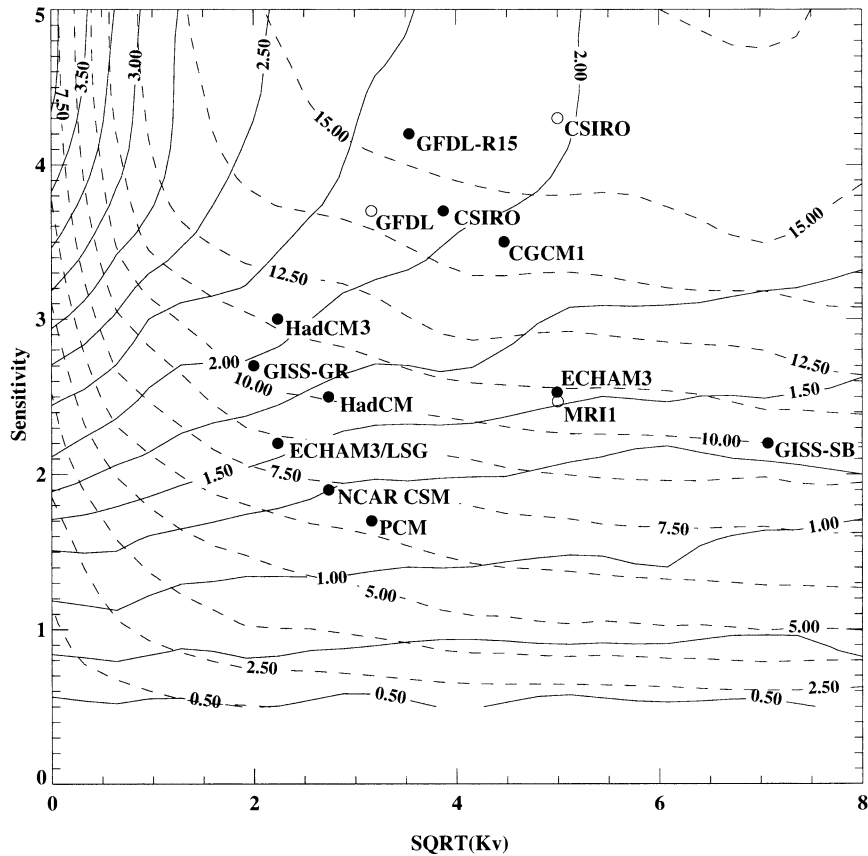
FIG. 1. Changes in surface air temperature (solid curves) and sea level rise due to thermal expansion of the ocean (dashed curves) at the time of CO_2 doubling. Positions of the fits for different AOGCMs using an effective climate sensitivity are shown by closed circles. Positions of the versions of the MIT model that reproduce changes in SAT for ECHAM3/LSG, CSIRO, and GFDL R15 using their equilibrium sensitivities are shown by open circles.

models. In both cases they are somewhat different from those obtained in the simulations with corresponding AGCMs coupled to 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.[2] The energy balance of the climate system can be described by the following simple equation:

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

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, $\lambda_{eq} = F_{2\times CO_2}/\Delta T_{eq}$, where $F_{2\times CO_2}$ is a forcing due to CO_2 doubling. In a transient run, a time-dependent effective feedback parameter can be estimated as follows:

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

where $R_{toa}(t)$ is the net radiative flux at the top of the atmosphere. An effective climate sensitivity, ΔT_{eff}, is then defined as what the equilibrium surface warming due to CO_2 doubling would be if $\lambda = \lambda_{eff}$, $\Delta T_{eff} = F_{2\times CO_2}/\lambda_{eff}$. The values of the effective sensitivity at the time of CO_2 doubling for some AOGCMs used in the

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 $2 \times CO_2$
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

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[2] Changes in the model sensitivity described by Senior and Mitchell (2000) occurring after several 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.
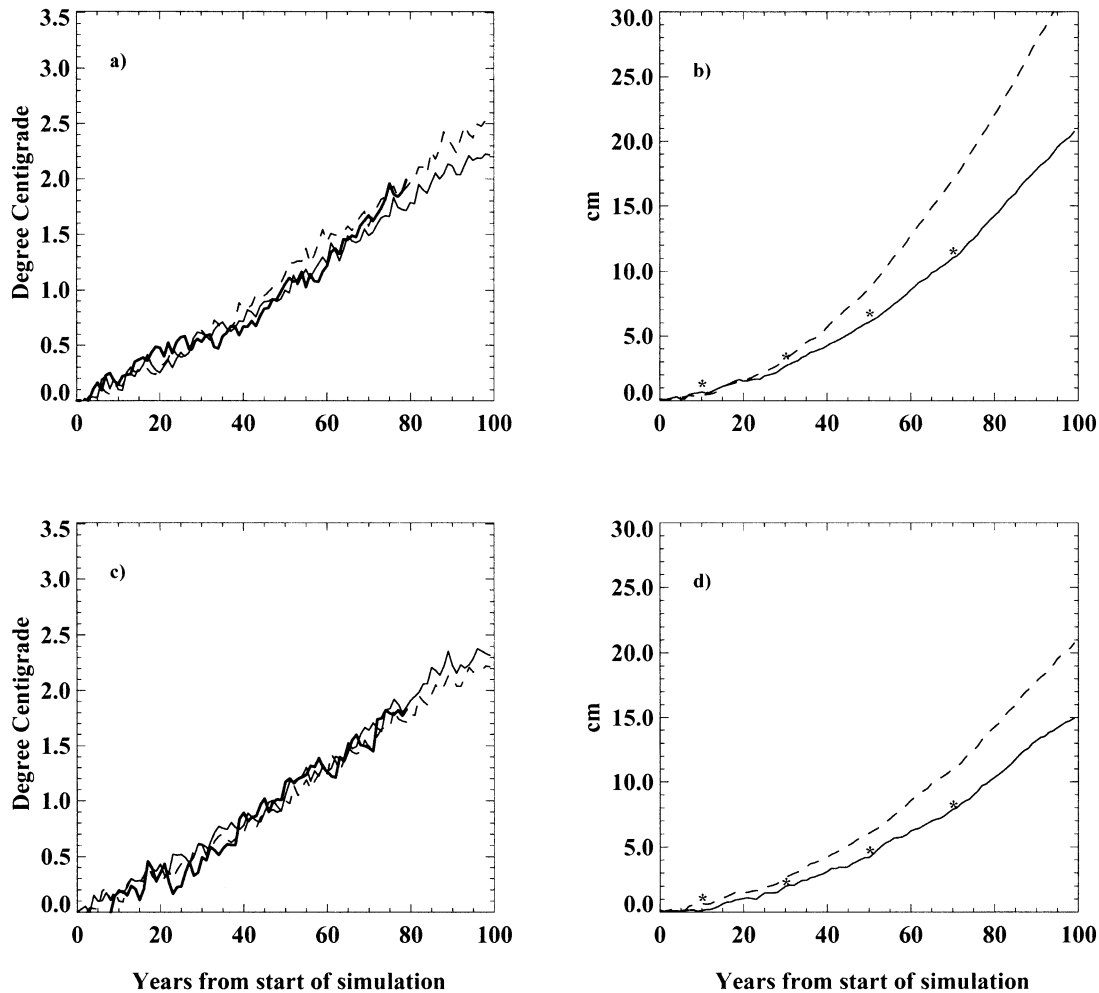
FIG. 2. Changes of annual mean global mean surface air temperature and sea level (thermal expansion) in simulations with the (a), (b) MRI1 and (c), (d) ECHAM3/LSG AOGCMs and in simulations with the versions of the MIT 2D climate model with effective (thin solid lines) and equilibrium (dashed lines) climate sensitivities. Data from CMIP2 simulations with AOGCMs are shown by thick solid line (SAT) and by asterisks (sea level). 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-yr mean for four consecutive segments of the simulations. In this study we used data on sea level rise for these four periods provided by S. Raper (Raper et al. 2002).

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.

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 though they fit the SAT changes. For example, very large effective diffusion coefficients ($K_v = 500$ $cm^2 \, s^{-1}$) are required to reproduce changes in SAT simulated by the Meteorological Research Institute 1 (MRI1) AOGCM (Fig. 2a) when the model's equilibrium climate sensitivity of 4.8°C is used. However, the

MIT climate model with these parameters produces a significantly larger sea level rise (Fig. 2b). At the same time, the MIT model with S matching the effective sensitivity of the MRI1 AOGCM (i.e., $S = 2.6$°C) and K_v $= 50 \, cm^2 \, s^{-1}$ reproduces changes in both SAT and sea level.

Using the effective sensitivity, instead of an equilibrium one, leads to a 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 Commonwealth Scientific and Industrial Research Organisation (CSIRO) and GFDL R15 models (Fig. 3). This difference between simulations with different mod-
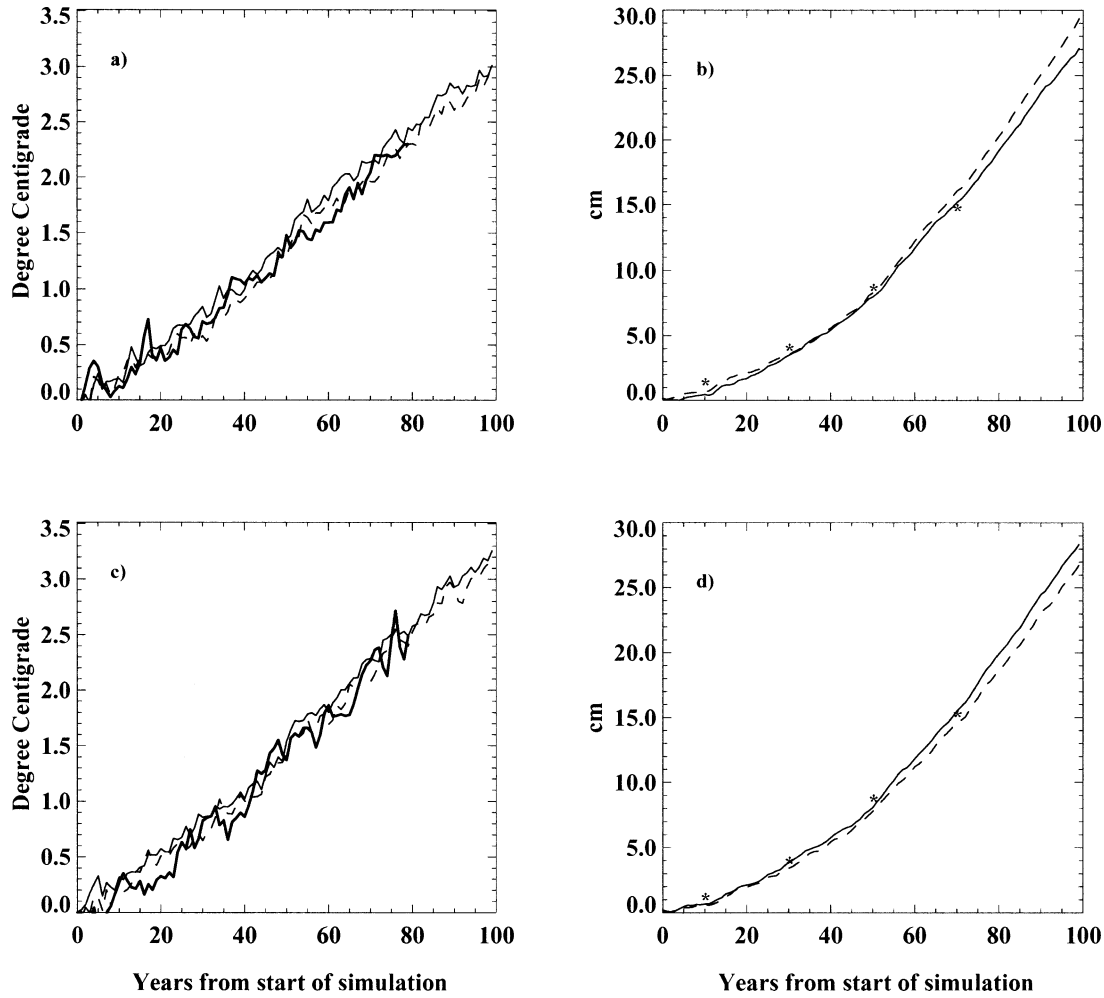
FIG. 3. As in Fig. 2 but for the (a), (b) CSIRO and (c), (d) GFDL R15 AOGCMs.

els is explained by different sensitivities of SAT and sea level to change in K_v for different values of S. The latter can be explained through simple analysis of Eq. (1). For linear forcing $F(t) = \gamma t$, Eq. (1) has an analytical solution under the assumption that C is fixed; namely,

$$\Delta T_s(t) = \gamma S'[t - \tau(1 - e^{t/\tau})], \qquad (3)$$

where $S' = \lambda^{-1}$ is the "specific" sensitivity (i.e., an equilibrium SAT increase due to forcing of 1 W m^{-2}) and $\tau = S'C$.

However, for Eq. (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 Eq. (3) is not an exact solution of Eq. (1) for time dependent C, it approximates a numerical solution of Eq. (1) rather well with τ proportional to $S'\sqrt{K_v t}$.

The dependence of SAT increase on the rate of heat mixing into the deep ocean is defined by the e-folding time constant τ, which is proportional to climate sensitivity. For small τ, $\Delta T_S(t)$ becomes proportional to $\gamma S' t$ which is independent of K_v. In other words, for low values of S, changes in SAT at any given time are close to the equilibrium response to a corresponding forcing regardless of how much heat is taken down. For large τ, $\Delta T_S(t) \approx \gamma S' t^2/\tau = \gamma t^2/C$ which is proportional to $(K_v)^{-1/2}$. This explains the different sensitivity of SAT to changes in K_v for different values of S. The difference in sensitivity of sea level rise to change in effective diffusion coefficient for high and low climate sensitivity is explained by the difference in SAT's sensitivity to K_v. An increase in K_v leads to an increase in ocean model capacity to mix mixed layer temperature anomalies into the deep ocean. This leads to an increase in thermal expansion. For high climate sensitivities this effect is partly offset by a decrease in mixed layer temperatures in response to an increase in K_v. This does not happen in the case of low climate sensitivities making sea level relatively more sensitive to changes in K_v.

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

Model	Forcing (W m^{-2})
CSIRO	3.45
HadCM2	3.47
HadCM3	3.74
NCAR CSM	3.60
PCM	3.60
GISS and MIT 2D	3.84

For example, an increase in effective diffusion coefficient from 0.5 to 12.5 cm^2 s^{-1} leads to a doubling of sea level rise for $S = 1.6°C$ while causing just a 40% change for $S = 4.5°C$ (Fig. 1; see also Sokolov and Stone 1998). It should be noted that an absolute change in thermal expansion caused by the same change in K_v is larger for high climate sensitivity.

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 heat uptake (Fig. 1). Sea level rise, in contrast with SAT, is rather sensitive to changes in K_v in this region of the parameter space. This, together with the relatively small increase in sea level projected by the ECHAM3/LSG model, explains the noticeable difference between sea level rise produced by this model's fits with equilibrium and effective sensitivities. The opposite is true for both the CSIRO and the GFDL R15 models. For high climate sensitivity SAT is rather sensitive to changes in K_v while sea level is not. Therefore, fits with different S for these models have similar K_v and produce similar sea level changes. It should be noted that the difference in sea level rise projections based on 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 for all models. (We recall that fits with both effective and equilibrium S were constructed so as to produce similar changes in SAT.)

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 fol-

[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 to reproduce the behavior of AOGCMs (Cubasch et al. 2001; Raper et al. 2002).

lowing way. As is well known, radiative forcing for an exponentially increasing $CO_2(t) = CO_2(0)*\exp(\alpha t)$ can be written as $F(t) = \kappa \ln[CO_2(t)/CO_2(0)] = \kappa \alpha t = \gamma t$, where α is a rate of CO_2 increase and κ is a coefficient different for different models, and $\gamma = \kappa \alpha$. A value of κ for a given model is defined by the details of its radiation code (for the MIT 2D model $\kappa = 5.35$ W m^{-2}) 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 of a given AOGCM rather than its sensitivity to CO_2 doubling. For example, due to differences in the forcing corresponding to the CO_2 doubling (3.84 W m^{-2} vs 3.45 W m^{-2}), $S = 3.7°C$, matching the equilibrium sensitivity of the CSIRO AOGCM, corresponds to a warming of $0.96°C$ (W m^{-2})$^{-1}$ for the MIT 2D model while the "specific" sensitivity of the CSIRO AOGCM is $1.07°C$ (W m^{-2})$^{-1}$. Therefore, a climate sensitivity of $4.12°C$ should be used in the simulation with the MIT 2D model to match the specific sensitivity of the CSIRO AOGCM. In general, for the models with a forcing smaller than produced by our model, we need to use S higher than their effective sensitivities and vice versa.

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 (Fig. 4) show small differences compared to the simulations with the original sensitivities and forcing. Such a small impact of different forcings on the results of simulations with increasing CO_2 can be explained using Eq. (3). 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 for 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. This might be due in part to the relatively small differences in forcing between different models, less than 10%. Because data on radiative forcing are not available for all models, the estimates for K_v 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 [modified version of the model described in 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 model sensitivities. Fits for some models used in CMIP2 simulations were not obtained due to absence of data on sea level rise.

Because, as discussed above, the time delay constant τ is proportional to $\sqrt{K_v}$, $\sqrt{K_v}$ seems to be the most natural measure for the rate of oceanic heat uptake for
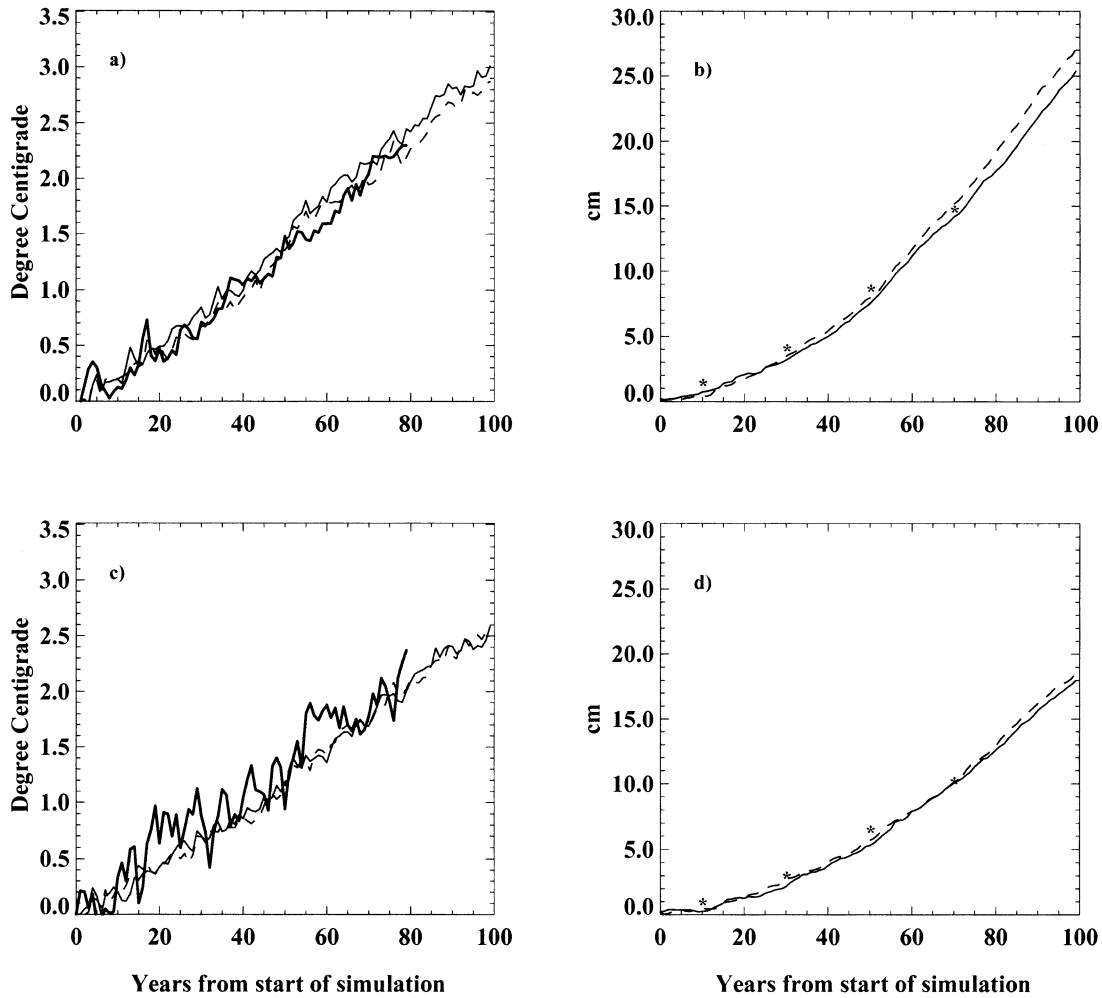
FIG. 4. Changes of annual mean global mean surface air temperature and sea level (thermal expansion) in simulations with the (a), (b) CSIRO and (c), (d) HadCM2 AOGCMs and in simulations with the versions of the MIT 2D climate model with corrected (dashed lines) and uncorrected (thin solid lines) forcing. Data from CMIP2 simulations with AOGCMs are shown by thick solid line (SAT) and by asterisks (sea level).

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.

Models	Parameters of corresponding versions of the 2D model		
	S (°C)	K_v (cm² s⁻¹)	$\sqrt{K_v}$ (cm s⁻¹/²)
CGCM1	3.6	20	4.47
CSIRO	3.7	15	3.87
ECHAM3/LSG	2.2	5	2.24
GFDL R15	4.2	12.5	3.54
GISS GR	2.7	4.0	2.0
GISS SH	2.2	25.0	5
HadCM2	2.5	7.5	2.74
HadCM3	3.0	5.0	2.24
MRI1	2.6	25.0	5.0
NCAR CSM	1.9	7.5	2.74
PCM	1.7	10.0	3.16

the MIT 2D model. For models given in Table 3, $\sqrt{K_v}$ varies from 2.0 to 5.0 cm s⁻¹/². In Fig. 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. 2002). The PDF for models was obtained by fitting an analytical distribution to data from Table 3. Data for all models were weighted equally. In Forest et al. (2002) a joint PDF for K_v, S, and net aerosol forcing was obtained by estimating the goodness-of-fit between results of the MIT model's simulations and observations, while accounting for internal variability of the climate system. The marginal PDF for K_v is shown in Fig. 5. Though shapes of the two PDFs are different, values of K_v for all models fall into the 5%–95% confidence interval calculated from observations. The means (medians) of the two distributions are also not very different, 3.21 (3.01) cm s⁻¹/² and 4.20 (4.40) cm s⁻¹/² for models and observations, respectively. It should be noted that, be-
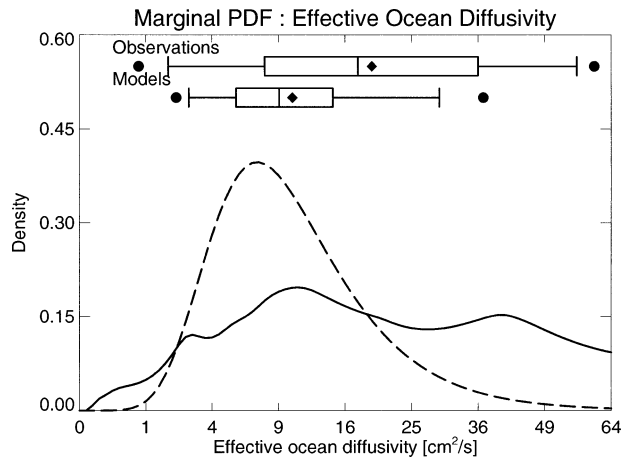
FIG. 5. Probability density functions for the rate of oceanic heat uptake from models (dashed) and observations (solid). The whisker plots show the 2.5%–97.5% (dots), 5%–95% (vertical bars on ends), and 25%–75% (box) probability ranges along with the median (bar within box) and mean (diamond) for each distribution.

cause the observations do not place an upper bound on K_v, a subjective bound of $K_v = 64$ cm^2 s^{-1} was imposed. For a different choice of an upper bound the confidence interval would be somewhat different.

4. 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 average 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 a factor of 3 (in terms of $\sqrt{K_v}$). This difference would by itself introduce considerable uncertainty in long-term projections of climate change. We note that the values for all models fall within the 5%–95% interval of the range derived from comparisons with the 20th century climate record (Forest et al. 2002). However, the range of K_v values derived from models is significantly narrower than that suggested by observations. Thus estimates of uncertainty that are based solely

on the model range like those of the Intergovernmental Panel on Climate Change (Cubasch et al. 2001) underestimate the true uncertainty in the rate of oceanic heat uptake.

APPENDIX

Change in Precipitation

As was 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.

Changes in precipitation, however, 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 (Figs. A1a and A1b), but significantly overestimates the increase in precipitation for the CGCM1 model and underestimates it for the MRI1 model (Figs. A1c and A1d). Figure A2 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. Covery 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 aforementioned strong correlation between changes in SAT and precipitation for the MIT model simulations is not surprising. As a result, the MIT climate model does not simulate climate changes that are cold and wet or hot and dry. Nevertheless, it reproduces the range of increases in precipitation produced by AOGCMs.

The aforementioned correlation can also be found in simulations with other models. For example, equilibrium $2 \times CO_2$ simulations with different version of the
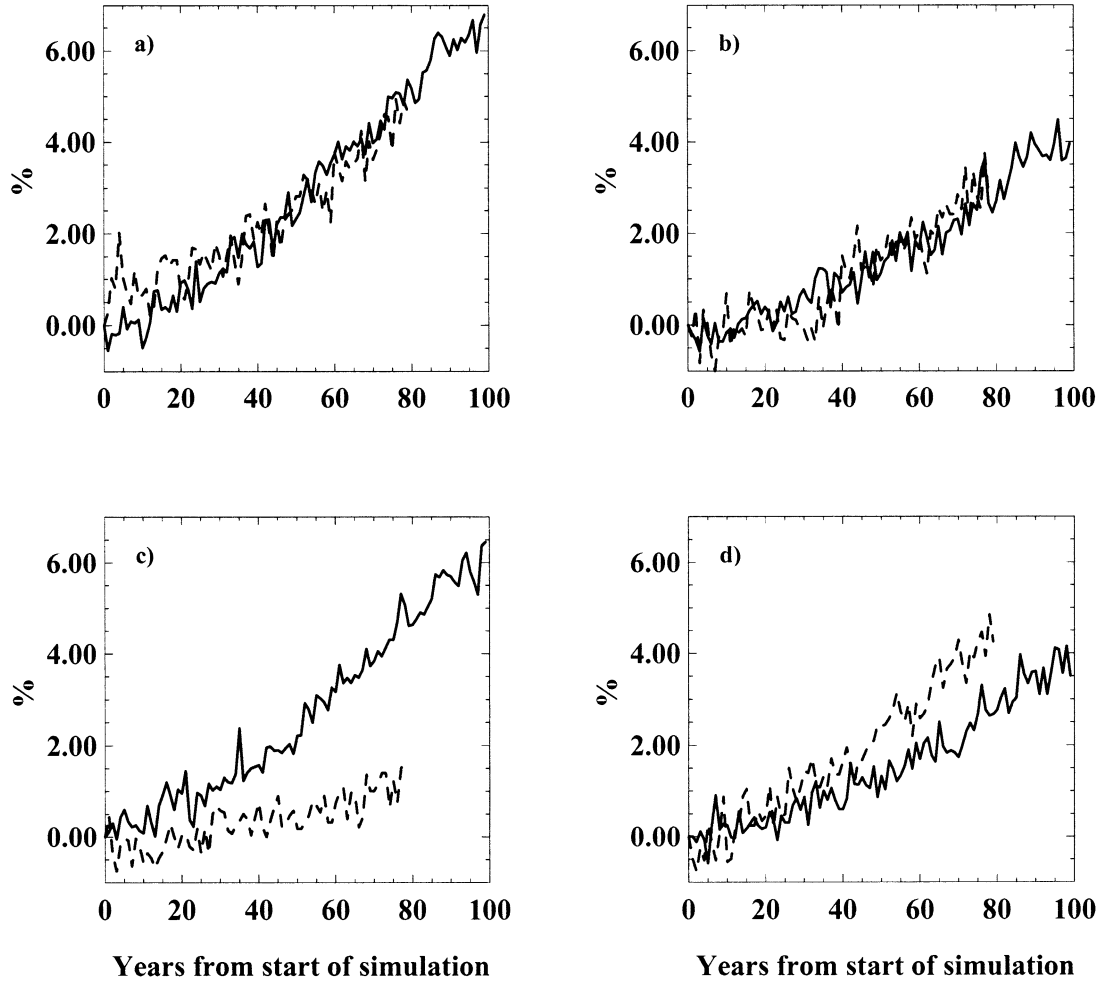
FIG. A1. Changes of annual mean global mean precipitation in simulations with (a) CSIRO, (b) ECHAM3, (c) MRI, and (d) CGCM1 AOGCMs (solid lines) and with the matching versions of the MIT 2D climate model (dashed lines).
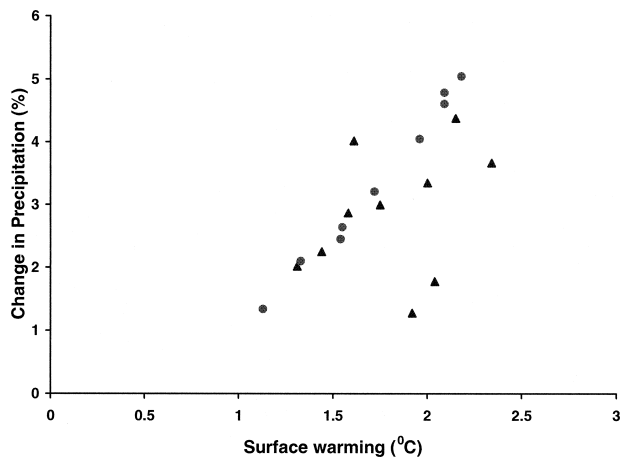

FIG. A2. Percentage change in globally and annually averaged precipitation as a function of global mean warming at the time of doubling of CO_2 as produced by different AOGCMs (triangles) from CMIP2 and corresponding versions of the MIT 2D model (circles).

UKMO model show a very similar linear dependence of changes in global precipitation on changes in SAT [see Fig. 9a in Sokolov and Stone (1998)]. The only difference between those versions was cloud parameterization.

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