A Flexible Climate Model For Use In Integrated Assessments

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Abstract

Because of significant uncertainty in the behavior of the climate system, evaluations of the possible impact of an increase in greenhouse gas concentrations in the atmosphere require a large number of long term climate simulations. Studies of this kind are impossible to carry out with coupled atmosphere ocean general circulation models (AOGCMs) because of their tremendous computer resource requirements. Here we describe a two-dimensional (2D, zonally averaged) atmospheric model coupled with a diffusive ocean model developed for use in the integrated framework of the MIT Joint Program on the Science and Policy of Global Change. The 2D model has been developed from the GISS GCM and includes parameterizations of all the main physical processes. This allows it to reproduce many of the nonlinear interactions occurring in simulations with GCMs. Comparisons of the results of present-day climate simulations with observations show that the model reasonably reproduces the main features of the zonally averaged atmospheric structure and circulation.

The model's sensitivity can be varied by changing the magnitude of an inserted additional cloud cover feedback. Equilibrium responses of different versions of the 2D model to an instantaneous doubling of atmospheric CO_2 are compared with results of similar simulations with different AGCMs. It is shown that the additional cloud feedback does not lead to any physically inconsistent results. On the contrary, changes in climate variables such as precipitation and evaporation, and their dependencies on surface warming produced by different versions of the MIT 2D model are similar to those shown by GCMs.

By choosing appropriate values of the deep ocean diffusion coefficients, the transient behavior of different AOGCMs can be matched in simulations with the 2D model, with a unique choice of diffusion coefficients allowing one to match the performance of a given AOGCM for a variety of transient forcing scenarios. Both surface warming and sea level rise due to thermal expansion of the deep ocean in response to a gradually increasing forcing are reasonably reproduced on time scales of 100–150 years. However a wide range of diffusion coefficients is needed to match the behavior of different AOGCMs. We use results of simulations with the 2D model to show that the impact on climate change of the implied uncertainty in the rate of heat penetration into the deep ocean is comparable with that of other significant uncertainties.

1. Introduction

The climatological impact of increases in greenhouse gas concentrations in the atmosphere, despite being a subject of intensive study in recent years, is still very uncertain. Simulations with coupled atmosphere-ocean general circulation models (AOGCMs) produce significantly different climate changes in response to the increase of greenhouse gas (GHG) concentrations in the atmosphere. There are a number of reasons for these differences. Climate sensitivities, that is equilibrium increases of surface air temperature in response to the doubling of the CO_2 concentration, range from 1.9 to 5.4 ˚C, for different models (Senior and Mitchell 1993, IPCC 1996). Another uncertainty affecting possible climate change is differences in heat uptake by the deep ocean in different ocean models. There are, also, significant uncertainties in the projected increase of the concentrations of GHGs and aerosols and in the corresponding radiative forcing.

One of the main goals of the MIT Joint Program on the Science and Policy of Global Change is to evaluate the role of these uncertainties and their interactions in climate change predictions

(Jacoby and Prinn, 1994). Only very limited studies of this kind can be performed with AOGCMs due to their large requirements for computational resources. Moreover, the only uncertainty that can be addressed in simulations with a given AOGCM is uncertainty in the forcing (Cubasch, *et al.,* 1992; IPCC, 1996). In most cases different upwelling diffusion-energy balance (UD/EB) models have been used for studying uncertainty in climate change (IPCC, 1990, 1992 and 1996; Murphy, 1995; Wigley and Raper, 1993; Jonas, *et al.*, 1996).

A modified version of the 2D (zonally averaged) statistical-dynamical atmospheric model (Yao and Stone, 1987; Stone and Yao, 1987 and 1990) developed on the basis of the GISS GCM (Hansen, *et al*., 1983) has been chosen for use in climate simulations in the integrated framework of the MIT Joint Program. Unlike energy balance models, the 2D model includes parameterizations of all the main physical processes and is, therefore, capable of reproducing many of the nonlinear interactions taking place in GCMs. At the same time it is about 20 times faster than the GISS GCM with the same latitudinal and vertical resolutions. At the present time the 2D model is coupled with a simple diffusive ocean model. In climate change simulations discussed below, the 2D climate model is driven by changing greenhouse gas concentrations. Studies of climate change caused by different emission scenarios (Prinn, *et al.*, 1996) have also been carried out with a version of the MIT 2D climate model which includes fully interactive atmospheric chemistry and transport of chemical species (Wang, *et al*., 1995), and calculates oceanic carbon uptake.

Brief descriptions of both the atmospheric and oceanic models are given in Section 2, with some results of a present-day climate simulation presented in Section 3. For studying uncertainties in climate change, model versions with different sensitivities were obtained by changing the cloud feedback in a way described by Hansen, *et al.* (1993). In Section 4 responses of different versions of the MIT 2D model to the doubling of the CO_2 concentration are compared with the results obtained in simulations with different AGCMs. This comparison shows that the 2D model's behavior is similar to that of AGCMs. It is also shown that the 2D model can match the transient responses of different AOGCMs to a gradual increase in atmospheric CO_2 by using different diffusion coefficients. Uncertainties in projected climate change, namely, in surface warming and sea level rise, associated with uncertainties in the climate sensitivity and rate of heat uptake by the deep ocean are studied in the Section 5. Finally, some conclusions are given in Section 6.

2. Model Description

A detailed description of the original GISS 2D model is given in Yao and Stone (1987) and Stone and Yao (1987 and 1990). The model solves the primitive equations as an initial value problem, using a finite-different approximation in latitude-pressure coordinates. The grid used in the model contains 24 points in latitude, corresponding to a resolution of 7.826˚. The model has nine layers in the vertical: two in the planetary boundary layer, five in the troposphere, and two in the stratosphere. The model's numerics and most of the parameterizations of physical processes (radiation, convection, *etc*.) are closely parallel to those of the GISS GCM (Hansen *et al.,* 1983). A very important feature of the model, from the point of view of both climate change study and coupled chemistry-climate dynamics, is the radiation code of the GISS GCM that it incorporates. This code includes all significant greenhouse gases, such as H_2O, CO_2, CH_4, N_2O, CFC's, *etc*., and 11 types of aerosols.

A number of modifications have been made to the model at MIT to make it more suitable for climate change studies (Sokolov and Stone, 1995; Prinn, *et al.*, 1996; Xiao, *et al.*, 1997). The first one was to include in the 2D model a real land-ocean distribution. The modified MIT 2D model, as well as the GISS GCM, allows up to four different kinds of surface in the same grid cell, namely, open ocean, ocean-ice, land, and land-ice. The surface characteristics (*e.g.*, temperature, soil moisture, albedo) as well as turbulent and radiative fluxes are calculated separately for each kind of surface while the atmosphere above is assumed to be well mixed horizontally. The area weighted averages of fluxes from different kinds of surfaces are used to calculate the change of temperature, humidity, and wind speed in the atmosphere.

In the GISS GCM turbulent fluxes of temperature, moisture and momentum are calculated under the assumption that the atmospheric surface layer is in equilibrium, which leads to a complicated algorithm including nested iterations. That algorithm, while used successfully in both the GISS GCM and 2D model without land, produces computational problems when land is included. Because of this, the equilibrium assumption has been replaced by the assumption that the layer between the surface and the model's first level is well mixed. The surface wind speed calculation also has been simplified compared to the procedure used in the GISS GCM. The absolute value of $\left|\vec{V}_S\right|$ is assumed to be equal to:

$$\left|\vec{V}_S\right| = \sqrt{\vec{V}_1^{\,2} + \overline{\left(V'^{2}_{s}\right)}},$$

where $\vec{V}_1$ is wind at the model's first level and $\overline{\left(V'^{2}_{s}\right)}$ is the zonal mean surface wind variance. Calculation of the surface wind variance is similar to that of temperature variance described in Yao and Stone (1987). The cross isobar angle, α, is calculated as a function of $\left|\vec{V}_S\right|$ and a bulk Richardson number Ri_s, according to formulas derived by G. Russell (personal comm., 1993):

$$\alpha = \frac{0.0625 \times 2\pi}{1 + \left|\vec{V}_S\right|\alpha_0}, \text{ if } Ri_s < 0, \text{ and}$$

$$\alpha = \frac{2\pi\left(0.09375 - \dfrac{0.03125}{1 + 4Ri_S}\right)}{1 + \left|\vec{V}_S\right|\alpha_0}, \text{ if } Ri_s > 0.$$

Here α_0 is an empirical coefficient that relates cross isobar angle to the surface wind which is taken equal to 0.3.

Two different types of clouds are taken into account in the model: convective clouds, associated with moist convection, and large-scale or supersaturation clouds, formed due to large-scale condensation. The amount of convective clouds in a given layer is proportional to the mass flux due to moist convection through the lower boundary of this layer. The amount of supersaturated clouds is expressed as a height-dependent function of the critical value of relative humidity for cloud formation and the critical value of relative humidity for condensation (h_{con}). Due to course model resolution, a value of h_{con} = 90% has been chosen as the criterion for condensation. Convection occurs in just the unstable fraction of a latitude belt, based on a parameterization of subgridscale variations of temperature and humidity (Yao and Stone, 1987).

Some changes have been made to the GISS 2D model's radiation calculations. A dependence of snow albedo on surface temperature has been included; snow density and heat conductivity are now calculated as functions of snow mass instead of being fixed, and the temperature distribution in the ground layer is assumed to be linear rather than quadratic as in the GISS GCM.

In the GISS 2D model a 16^{th}-order filter is applied to air temperature and specific humidity. We found that this has a significant impact on the present-day climate simulations and the model's climate sensitivity. For example, the zonal wind simulated by the 2D model without filter is much closer to the observed, especially for Northern Hemisphere winter. On the other hand, specific humidity in the equatorial region turned out to be unrealistically low. The latter was improved by incorporating into the model a second order horizontal diffusion for specific humidity. This diffusion is, however, small compared to the model's parameterised diffusion due to large-scale eddies outside the tropics. Results of simulations with both the MIT 2D model and chemistry-transport model (Wang, *et al.*, 1995), as well as with the original GISS 2D model (Stone and Yao, 1990), have shown that the original 2D model's parameterisation of eddy transports tends to underestimate transports of heat and tracers. Consequently the value of the scaling constant A in Equation 1 in Stone and Yao (1990) was increased from 0.6 to 0.9.

In simulations of the present-day climate and equilibrium climate change the 2D model is coupled with a mixed-layer ocean model. In order to simulate the current climate, the equation for the mixed-layer temperature includes a term representing the effect of horizontal heat transport in the ocean and heat exchange between the mixed layer and deep ocean. The heat balance equation for the mixed layer is expressed in terms of the specific heat capacity and density of salt water, sea-ice mass, mixed layer depth, latent heat of freezing, heat balance on the ocean surface, heat flux through the lower surface of sea-ice, and fractions of grid cell covered by open ocean and sea-ice. The horizontal heat flux can be calculated from this equation using the results of a climate simulation with prescribed climatological sea surface temperature and sea-ice distributions. This ocean treatment is essentially similar to the one used by Meleshko, *et al*. (1991), except that in the MIT model the mixed layer depth is a prescribed function of season and latitude (Hansen et al., 1988). The algorithm used for calculation of the thermal energy of the mixed layer with variable depth is described in Russell, *et al.* (1985). The above mentioned change in the eddy flux parameterization led to a better agreement of the implied ocean heat transport with observations.

In simulations of transient climate change the heat uptake by the deep ocean has been parameterized by diffusive mixing of the perturbations of the temperature of the mixed layer into deeper layers (Hansen, *et al.*, 1988). The zonally averaged values of diffusion coefficients calculated from measurements of tritium mixing have been chosen as "standard" ones (Table 2.1).

Table 2.1. Coefficients of heat diffusion into the deep ocean (cm^2/s).

Northern Hemisphere											
90˚ N	82˚N	74˚N	66˚N	59˚N	51˚N	43˚N	35˚N	27˚N	20˚N	12˚N	4˚N
0.76	1.44	3.31	4.63	5.14	3.57	2.57	1.62	1.34	0.54	0.22	0.23
Southern Hemisphere											
4˚S	12˚S	20˚S	27˚S	35˚S	43˚S	51˚S	59˚S	66˚S	74˚S	82˚S	90˚S
0.32	0.43	1.24	1.53	2.61	4.67	6.97	7.60	8.11	9.73	0.00	0.00

The global average value of the diffusion coefficients, denoted as K_v, equals 2.5 cm^2/s for these "standard" values. However, Hansen, *et al.* (1984) found that the equivalent value of diffusion coefficient that gives similar results when used in a 1D model is only 1 cm^2/s. As will be shown in Section 4, two times the "standard" diffusion is required to match the behavior of the UD/EB model used in IPCC, 1996, which uses a diffusion coefficient equal to 1 cm^2/s, but also includes upwelling with a rate which decreases as global temperature increases.

3. Present-Day Climate Simulation

In developing the MIT 2D climate model a significant number of present-day climate simulations were performed and different versions of the above mentioned parameterization schemes were tested (Sokolov and Stone, 1995). Here we show the results averaged for twenty years of a run carried out with the 2D model coupled with the mixed layer ocean model as described above. Figures 3.1 and 3.2 show zonal wind and meridional streamfunction averaged for December, January, February (DJF) and June, July, August (JJA) for the 2D model, both of which are simulations similar to those observed (*e.g.*, see Peixoto and Oort, 1992), except for the streamfunction being too weak in Northern Hemisphere winter. Precipitation and evaporation obtained in simulations with the 2D model and the GISS (Model II) GCM (Hansen, *et al.*, 1983) are shown in Figures 3.3. and 3.4 together with observations (Leemans and Cramer, 1990; Oberhuber, 1988). Both the 2D and GISS models have difficulty matching the observed precipitation particularly, in DJF (Figure 3.3). The underestimation of precipitation in the tropics in DJF by the 2D model is consistent with the above mentioned deficiency of the simulated mean meridional circulation. As one would expect, the agreement between the results of the 2D model simulation and observed precipitation in the equatorial region is much better in JJA. It should be mentioned that there are in any case significant disagreements among observational data sets for precipitation. The pattern of evaporation is reasonably well reproduced by the 2D model.

Neither the 2D model nor the GISS GCM reproduces the seasonal cloud change (see Sokolov and Stone, 1995) in the tropics associated with the shift of the Intertropical Convergence Zone. Nevertheless, the overall pattern of seasonal changes in clouds (see Sokolov and Stone, 1995) and cloud radiative forcing (Figure 3.5) simulated by the 2D model, is quite similar to the observed.

As mentioned above, the eddy flux parameterization has been slightly changed, so as to increase eddy transports. As a result, total atmospheric energy transport (Figure 3.6) is in better agreement with observation (*e.g.*, Trenberth and Solomon, 1994) than it was in a previous version of the MIT 2D model (Sokolov and Stone, 1995). The same is true for the implied ocean heat transport, which is, however, still higher than observed (*e.g.*, Trenberth and Solomon, 1994), especially, in the Northern Hemisphere tropics, because of the model's underestimation of the Hadley cell heat transport in the Northern Hemisphere.

There is one essential problem with the simulation of sea surface temperature (SST) and sea ice distribution in a 2D model. Because of longitudinal variations of sea surface temperature and sea ice, zonal mean SST may be above 0 ˚C even when part of the ocean surface at the same latitude is covered by ice. However, in the formulation of the mixed layer ocean model, SST is kept at or below 0 ˚C until all ice melts, and no sea ice forms if SST is above the freezing point for salt water, that is –1.56 ˚C. As a result, the above mentioned feature of the SST and sea ice distribution cannot be simulated by a 2D model. Because of this, data used in the simulations with prescribed SST and sea ice have been adjusted. Namely, if, in any given latitude belt, less than 10% of the ocean surface is covered by ice and zonal mean SST is above 0 ˚C ice is removed. If ice covers more than 10% of the ocean, then SST is set to 0 ˚C if the mass of ice is decreasing, and to –1.56 ˚C otherwise. The adjusted data are shown in Figures 3.7 and 3.8 instead of direct observations. As one can see, despite the use of a “Q-flux” in the mixed layer model there are some differences in the prescribed and simulated values of SST and sea ice cover and depth. Some reasons for these differences are described in Sokolov and Stone (1995). Similar problems with predicting sea ice distribution in simulations with the GISS GCM with $8^\circ \times 10^\circ$ resolution are discussed by Hansen, *et al.* (1984).

Variations of the globally averaged annual mean surface air temperature in two simulations of the present-day climate with the 2D model are shown in Figure 3.9. The standard deviations of surface temperature in the simulation with the 2D model coupled with just the mixed layer ocean model is 0.10 ˚C; however, it decreases to 0.06 ˚C if mixing of mixed layer temperature perturbations into the deep ocean is taken into account. Thus, the unforced interannual variability produced by the MIT 2D model is rather close to that of the GISS AGCM but somewhat less than in simulations with coupled AOGCMs. The corresponding number for the GISS AGCM coupled with mixed layer model and diffusive deep ocean is 0.05 ˚C (Hansen, *et al*., 1997); and for the coupled GFDL and MPI AOGCMs (Santer, *et al.*, 1995) 0.10 ˚C and 0.12 ˚C, respectivly.

As a whole, a comparison of the model’s results with the observational data shows that it reproduces reasonably well the major features of the present-day climate state. Of course, there are some essential 3D features of the atmospheric circulation that cannot be simulated by a 2D model. However the depiction of the zonally averaged circulation by the 2D model is not very different from that by 3D GCMs. Since the model is to be used for climate change prediction, it is noteworthy that the seasonal climate variations are also reproduced quite well. Use of a 2D model allows as to perform a significantly larger number of climate simulations than would be possible with an AOGCM. A 100 year simulations takes about 12 hours on DEC Alpha Station 250 with the model described above and three times more with couple climate-chemistry model.

4. The 2D Model Response to Instantaneous and Gradual Increases in the Atmospheric CO_2 Concentration

4a. Equilibrium response to a doubling of CO_2 concentration

When using a 2D model to study uncertainty in climate change, it is desirable to have a model capable not only of simulating the present-day climate but also of reproducing the climate change pattern obtained in simulations with different GCMs. In this section responses of different versions of the MIT 2D model to an instantaneous doubling of CO_2 concentration in the atmosphere are compared with the results of similar simulations with different GCMs. The atmospheric model coupled with the mixed layer ocean model was used in these simulations. The seasonally changing horizontal heat transport by the ocean was specified from the results of the climate simulation with climatological sea surface temperature and sea-ice distribution and was held fixed, thereby neglecting the impact of possible changes in ocean circulation. This assumption has been always considered to be a significant weakness of equilibrium climate change simulations with mixed layer ocean models. However, as discussed in IPCC (1996), the results of a simulation with the GFDL AOGCM suggest that the oceanic heat transport for an equilibrium doubled CO_2 climate is similar to that for the present-day climate (see IPCC, 1996, Section 6.2.4). In any case the GCM simulations we compare with were carried out under the same assumption.

As mentioned in the introduction, the model’s versions with different sensitivities were obtained by inserting an additional cloud feedback, in the way proposed by Hansen, *et al.* (1993). Namely, calculated cloud amount is multiplied by the factor $(1 + k\Delta T_s)$, where ΔT_s is the increase of the globally averaged surface air temperature with respect to its value in the present-day climate simulation. Equilibrium responses of the MIT 2D model to the doubling of the atmospheric CO_2 for several values of k are shown in Table 4.1. The natural sensitivity, ΔT_{eq}, of the MIT 2D model, that is without an additional cloud feedback (k = 0), is 3.0 ˚C, about 1 ˚C less than for the version described in Sokolov and Stone (1995). This decrease in the model's sensitivity is mainly due to changes in the simulated water vapor distribution and somewhat weaker sea ice-albedo feedback especially in the Northern Hemisphere. The natural cloud feedback of the 2D model is very weak, *i.e.*, in a simulation with fixed clouds ΔT_{eq} equals 2.9 ˚C. Thus the ratio of ΔT_{eq} in the simulation with calculated cloud to that in the simulation with fixed cloud is 1.03 compared to 1.1 for the previous version of the 2D model (Sokolov and Stone, 1995), 1.25 for the GFDL (Wetherald and Manabe, 1988) and 1.75 for GISS (Hansen, *et al.*, 1984) GCMs. The magnitude of cloud feedback for the GISS GCM was evaluated by means of calculations with a 1D model using results of simulations with the GISS GCM (Hansen, *et al.*, 1984).

Table 4.1 Change in globally averaged annual mean surface air temperature, precipitation, cloud cover and cloud forcing due to a doubling of the atmospheric CO_2 .

k	ΔT_{eq}(˚C)	ΔP(%)	ΔCLD(%)	ΔNCF (w/m^2)
0.0875	1.6	0.8	3.5	-3.75
0.05	2.0	2.3	2.1	-3.11
0.02	2.5	4.6	0.5	-2.27
0	3.0	6.7	-1.2	-1.47
-0.008	3.5	8.9	-2.5	-0.84
-0.01	3.7	9.9	-3.0	-0.40
-0.0125	4.0	10.9	-3.5	-0.09
-0.015	4.6	12.6	-4.5	0.57

The equilibrium surface air temperature increase due to a doubling of the CO_2 concentration predicted by different GCMs ranges from 1.9 to 5.4 ˚C. A significant part of this difference is related to the differences in cloud feedbacks (Cess, *et al.*, 1990; Senior and Mitchell, 1993; Washington and Meehl, 1993) produced by the different GCMs. That, in turn, is caused mainly by different treatments of cloud optical properties. The feedback associated with changes in the optical properties of clouds in the GCM experiments is, of course, rather different from that associated with the changes in cloud amount, used in our simulations. However, it is interesting to note, that the relationship between changes in the globally averaged annual mean net cloud forcing (ΔNCF) and surface air temperature (ΔT_{eq}) in the simulations with different versions of the 2D model is qualitatively similar to that in the simulations with different versions of the UKMO GCM (Senior and Mitchell, 1993). In the simulations described by Senior and Mitchell ΔNCF equals –1.04, 0.21, 0.73, and 2.05 (W/m^2) for ΔT_{eq} equal to 1.9, 2.8, 3.3 and 5.4 ˚C, respectively.

Different versions of the MIT 2D model reproduce well the relationship between surface warming and increase in precipitation obtained in simulations with AGCMs (Figure 4.1). In Table 4.2 changes in the components of the globally averaged annual mean surface heat budget obtained in the simulation with the version of the MIT 2D model with ΔT_{eq}= 4 ˚C are compared with the results of AGCMs with similar sensitivities. The 2D model produces a relatively large increase in solar radiation absorbed by the surface which is mainly compensated by a larger decrease in latent heat flux than in the AGCMs. Changes in longwave radiation and sensible heat flux lie in the range produced by the AGCMs. A comparison of the results of the 2D model with those from the GISS AGCM shows that the larger increase in absorbed solar radiation is caused mainly by a substantial decrease in the cloud amount (see Table 4.1) and sea ice cover.

Table 4.2 Change in globally and annually averaged terms of the surface energy budget due to a doubling of the CO_2 concentration.

	NCAR	GFDL	GISS	MIT 2D
ΔT_{eq}	4.0	4.0	4.2	4.0
ΔLE	-7.4	-7.1	-10.4	-11.0
ΔH	2.5	0.6	3.3	2.1
ΔS	1.4	1.5	2.9	4.7
ΔF	3.7	4.9	4.1	4.2

Here S and F are the short and longwave radiation at the surface respectively; LE and H are the turbulent fluxes of latent and sensible heat. Results for AGCMs are from Washington and Meehl (1993).

At the same time the dependence of changes in the latent heat flux and radiation components of the surface energy budget on surface warming in simulations with different versions of the 2D model is very similar to that in simulations with different versions of the UKMO AGCM (Figures 4.2 and 4.3). In particular, although the 2D model's simulated change in solar radiation absorbed at the surface is quite different from that simulated by the NCAR, GFDL and GISS AGCMs (Table 4.2), it is consistent with the changes simulated by the UKMO AGCM (Figure 4.3). In contrast, the changes in sensible heat flux simulated by the 2D model are quite different from those simulated by the UKMO AGCM (Figure 4.2), but, as can be seen form Table 4.2, there is significant discrepancy in the changes of sensible heat flux even among GCMs with close sensitivities.

As discussed by Boer (1993), the CCC AGCM produces a decrease in the net solar radiation at the surface due to a negative feedback associated with an increase in cloud albedo. The versions of the MIT 2D model with sensitivities less than 2.5 ˚C and the version of the UKMO AGCM with calculated radiative properties of cloud show a similar change in solar radiation at the surface. The increase in evaporation/precipitation produced in the simulations with both the 2D model and the UKMO AGCM are, however, relatively larger than in the simulation with the CCC AGCM.

The latitudinal distributions of the CO_2 induced changes in surface air temperature, precipitation and cloud cover obtained in simulations with three versions of the MIT 2D model are shown in Figures 4.4 and 4.5 for DJF and JJA respectively. One of these versions is that without additional cloud feedback (k = 0). The other two are the versions that have sensitivities close to the high and low ends of the sensitivity range proposed by the IPCC (1990), namely 4.6 and 1.6 ˚C. Results of the doubled CO_2 simulation with the GISS GCM (Hansen, *et al.,* 1984) are also shown for comparison. All versions of the 2D model show an amplification of the surface warming in high latitudes, especially during winter. The GISS GCM produces an increase in surface temperature close to that of the 2D model with sensitivity 4.6 ˚C in the equatorial region and middle latitudes of the summer hemisphere and close to that of the 2D model with sensitivity 3.0 ˚C in high latitudes of the winter hemisphere. The sensitivity of the GISS GCM is 4.2 ˚C. At the same time, the latitudinal structure of surface air temperature change produced by the 2D model closely resembles the results of the GFDL and UKMO GCMs (Wetherald and Manabe, 1988; Senior and Mitchell, 1993; and Wilson and Mitchell, 1987).

Differences in the latitudinal gradient of surface warming between the 2D model and the GISS GCM can be explained by comparing changes in precipitable water and sea ice cover (Figures 4.6 and 4.7). The increase in precipitable water obtained in the simulation with the GISS GCM is larger in low latitudes than that produced by any version of the 2D model, and this causes a larger warming in the equatorial region. At the same time, the 2D model shows a significantly larger decrease in the sea ice cover in the winter hemisphere and, as a result, surface warming amplification, which is additionally increased by the decrease in cloud cover (Figures 4.4c, 4.5c and 4.10) in this region. The differences in the sea ice changes between the 2D model and the GISS GCM are, at least in part, due to the fact that the thickness of sea ice produced by the 2D model in the present-day climate simulation is less than that simulated by the GISS GCM (see Figures 3.7 and 3.8). As was shown by Rind, *et al*. (1995), both the global average and latitudinal profile of surface warming obtained in simulations with the GISS GCM depend significantly on the sea ice thickness used in the control simulation (which is not well constrained by observations); and on how the sea ice mass decrease caused by mixed layer warming is distributed between sea ice thinning and horizontal contraction. In the MIT 2D model, when mixed layer temperature reaches 0 ˚C additional heat is spent to reduce sea ice depth, until it reaches a minimal value, and is used to reduce horizontal extent after that, while in the simulation with the GISS GCM shown here it is spent equally on vertical and horizontal reduction of sea ice. There is, however, no difference between these two approaches after the sea ice depth reaches its minimal value.

The imposition of a minimal sea ice thickness leads to a discontinuity in the rate of sea ice cover decrease in response to global warming in a given grid cell. This, together with the low horizontal resolution of the 2D model, results in a discontinuity of the model sensitivity with respect to changes in the parameter k (see above given formula for cloud calculation). For $0.0875 \le k \le 0.0879$ $\Delta T_{eq} = 1.6$ ˚C, but ΔT_{eq} drops to 1.35 ˚C for k = 0.088. Such a sharp change is caused by the fact that in the last case about 30% of the ocean surface at 60 ˚S is covered by sea ice with minimal thickness, while sea ice is completely melted in the others. There is no such effect associated with sea ice changes in the high latitudes of the Northern Hemisphere, apparently, due to the smaller ocean area.

Height-latitude cross sections of the change in zonal mean temperature (Figures 4.8 and 4.9) show all the main features found in most GCM simulations with doubled CO_2, such as stratospheric cooling, maximum warming in the upper troposphere in the tropics and the above mentioned strong surface warming in high latitudes of the winter hemisphere. The magnitude of the decrease in stratospheric temperature is practically the same in all simulations regardless of sensitivity, similar to the results of the simulations with different versions of the UKMO AGCM (Senior and Mitchell, 1993). At the same time, the upper-tropospheric warming, in contrast to the results of the UKMO simulations (Senior and Mitchell, 1993 and Wilson and Mitchell, 1987), penetrates into the southern hemisphere in both DJF and JJA, more strongly during the northern hemisphere winter. Height-latitude cross sections of changes in annual mean zonally averaged cloud (Figure 4.10) bear an overall resemblance to the results produced by different GCMs (Hansen, *et al.*, 1984; Wetherald and Manabe, 1988; Senior and Mitchell, 1993). However, the increase of high cloud at low latitudes associated with the upward shift of the tropopause is almost absent in the simulation with the standard version of the 2D model (k = 0).

In general, the results presented above show that responses of different versions of the MIT 2D model to the doubling of the CO_2 concentration, in terms of both global average and zonal mean, are similar to those obtained in simulations with different GCMs. Since the climate model outputs are used in simulations with the Terrestrial Ecosystem Model (Prinn, *et al.*, 1996; Xiao, *et al.*, 1997), it is important to note that inserting the additional cloud feedback described above, while allowing us to change model sensitivity, does not lead to any physically unrealistic changes in climate. On the contrary, the changes in other climate variables, such as precipitation, evaporation and so on, are generally consistent with the results produced by different GCMs. It is worth noting that the ecosystem impacts of the climate change due to a CO_2 doubling, as simulated by the standard version of the MIT 2D model, are quite similar to those produced by the climate changes simulated by different AGCMs (Xiao, *et al.,* 1997).

4b. Transient response to a gradual CO_2 increase

The transient behavior of different AOGCMs can be matched by choosing appropriate values for the model's sensitivity and the rate of heat diffusion into the deep ocean. The change in the latter was obtained by multiplying the "standard" diffusion coefficients (Table 2.1) by the same factor at all latitudes, thereby preserving the latitudinal structure of heat uptake by the deep ocean. The time dependent globally averaged surface warming produced by different versions of the 2D model are compared with the results of simulations with the GFDL, MPI and NCAR AOGCMs in Table 4.3 and Figures 4.11 and 4.12. Values for sensitivities of the GCMs are taken from IPCC, 1996.

Table 4.3 Responses of different AOGCMs and matching versions of the MIT 2D model to a gradual increase of CO_2 concentration.

Model	ΔTeq (˚C)	ΔT at time of CO_2 doubling (˚C)	Fraction of equilibrium response (%)
GFDL	3.7	2.2	59
MIT 2D, K_v = 5	3.7	2.3	62
MPI	2.6	1.6	62
MIT 2D, K_v = 25.5	2.6	1.6	62
NCAR	4.6	3.8	83
MIT 2D, K_v = 0	4.6	3.9	84

As shown in Figure 4.11, the transient responses of the 2D model with the "standard" deep ocean diffusion coefficients doubled is very similar to those obtained in the simulations with the GFDL AOGCM with different rates of CO_2 increase (IPCC, 1996)[1]. Ten times "standard" values of the diffusion coefficients are required to match the delay in warming produced by the MPI

[1] In previous simulations (see Sokolov and Stone, 1996) we concluded that our "standard" diffusion coefficients give a good match to the GFDL AOGCM. However, the earlier simulations were carried out with a version of the 2D model with sensitivity 3.5 ^{0}C (the value for the GFDL GCM given by Murphy and Mitchell, 1995). In the simulations described here we use the value 3.7 ^{0}C for the GFDL model sensitivity, as given in IPCC, 1996, and by Stouffer (personal communication).

AOGCM (Cubasch, *et al.,* 1992) (see Figure 4.12).[2] Data for the MPI model have been corrected by taking into account error in the rate of surface warming associated with a “cold start” (Hasselmann, *et al.,* 1993; IPCC, 1996). At the same time, no heat diffusion into the deep ocean is required to reproduce the fast warming produced by the NCAR AOGCM (IPCC, 1996). The UD/EB model used in IPCC (1996) was tuned to reproduce the globally averaged results of the GFDL AOGCM. This implies, as noted in Section 2, that it has a rate of heat uptake close to that for the 2D model with doubled diffusion coefficients.

The only significant difference between results of the 2D model and the GFDL AOGCM occurs in the simulation with 0.25% per year increase in CO_2, and only after some 120–150 years of integration. Aside from that, the 2D climate model reproduces quite well the globally averaged surface warming predicted by different AOGCMs for a variety of forcing scenarios, for periods of at least 100 years.

At the same time, there is one noticeable difference in the zonal pattern of temperature increase obtained in the transient simulations with the MIT 2D model and that produced by most of the AOGCMs. That is, there is no strong interhemispheric asymmetry in the transient warming simulated by the 2D model. The change of zonally averaged surface air and sea surface temperature for the decade of doubling of CO_2 concentration obtained in the simulation with 1% per year increase in CO_2 with the version of the 2D model matching the GFDL AOGCM is shown in Figure 4.13, together with the equilibrium change obtained in the simulation with the same version couple with a mixed layer ocean model. The temporal evolution of the surface air temperature and sea surface temperature in the same transient simulation are shown in Figures 4.14 and 4.15, respectively. As can be seen from Figures 4.14 and 4.15, there is some delay in the surface warming in high latitudes of the southern hemisphere, much smaller, however, than in the similar simulation with the GFDL AOGCM (Manabe, *et al.*, 1991). The time dependent response of the sea surface temperature is closer to that in the simulation with the GFDL GCM, although the 2D model does not show cooling near Antarctica. The differences in the temporal changes of sea ice thickness (Figure 4.16) from that shown by Manabe, *et al.* (1991), are consistent with the differences in the surface warming. In particularly, in the simulation with the 2D model sea ice depth steadily decreases in both hemispheres as CO_2 increases, whereas it increases in the Southern Hemisphere in the GFDL simulation.

The deep ocean temperature change for the decade of CO_2 doubling in the simulation with the 2D model (Figure 4.17), bears a general resemblance with the results of the simulations with the GFDL AOGCM (Manabe, *et al.,* 1991) and with the UKMO AOGCM (Murphy and Mitchell, 1995). The 2D model produces zones of heat penetration into the deep ocean at about 60 ˚N and about 50 ˚S. Consistent with the significant surface warming in high latitudes of the Southern Hemisphere, the zone of very deep penetration of heat south of 60 ˚S is missing in the simulation with the MIT 2D model. However, some recent studies show that current ocean GCMs may produce excessive vertical mixing in this region and that, as a result, the corresponding retardation of warming may be exaggerated (IPCC, 1996).

Another characteristic describing changes in the deep ocean temperature is sea level rise due to thermal expansion. The thermal expansion has been calculated from the deep ocean temperature increase using the method described in Gregory (1993). Levitus’ (1992) data have been used for the unperturbed state of the deep ocean. In spite of our model’s simplified representation of the

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[2] It should be noted that radiative forcing produced by the MPI AGCM in response to the CO_2 doubling is about 0.5 W/m^2 less than that produced by the GISS AGCM (Cess, *et al*., 1993) and the MIT 2D model.

deep ocean, it reproduces the thermal expansion of the deep ocean as simulated by the GFDL AOGCM quite well (Figure 4.18), except for the already noted differences in the last stage of the simulation with 0.25% per year increase in CO_2. There is less agreement between results of the simulations with the MIT 2D model and the MPI AOGCM (Figure 4.19), especially for scenario D case. A similar problem in trying to reproduce the results of the MPI AOGCM with an UD/EB model was reported by Raper and Cubasch (see IPCC, 1996, Section 6.3.1).

5. Uncertainty in the Rate of Heat and Carbon Uptake by the Deep Ocean and its Impact on the Transient Model Response to the Increase of Atmospheric Greenhouse Gas Concentrations

Uncertainty in the rate of heat uptake by the deep ocean has not been included in the projections of climate change made by the Intergovernmental Panel on Climate Change (IPCC, 1990 and 1996). For example, the IPCC (1996) projections of a warming of 1 ˚C to 3.5 ˚C by 2100 only take into account uncertainties in greenhouse gas emission scenarios and in climate sensitivity, while the rate of heat uptake was chosen to reproduce results of the GFDL AOGCM. The same is true for IPCC projections of possible sea level rise. Among the reasons for not taking uncertainty in the heat uptake into account is, apparently, the conclusion of Wigley and Raper (1993) that it is relatively unimportant. However, the results for different coupled AOGCMs given above, together with the absence of any direct measurements of heat uptake by the ocean, indicate that there is more uncertainty than Wigley and Raper assumed. Thus, as an example of the use of the MIT 2D model, in this section we present results of simulations with the MIT 2D model concerning sensitivity of surface warming and sea level change to the rate of ocean heat uptake. While being qualitatively similar to those presented by Wigley and Raper (1993), the results lead us to a different conclusion.

Estimates of the impact of uncertainty in any given parameter depend strongly on both the range for this particular parameter and “best guesses” for the others. The range for the rate of heat uptake by the deep ocean used in the simulations below has been based in part on values needed to match the transient warming produced by different AOGCMs. The “standard” values of the diffusion coefficients, with $K_v = 2.5\ cm^2/s$, obtained, as mentioned above, from observations of tritium mixing into the deep ocean have been chosen as a “best guess.” The heat uptake by the deep ocean in 100 years produced by the version of the MIT 2D model using these coefficients is about 15% less than that when using the coefficients matching the GFDL AOGCM. The coefficients with $K_v = 12.5\ cm^2/s$, that is half as large as those matching the MPI AOGCM, have been used as a high end of the range. Since the zero rate of heat penetration into the deep ocean required to match the behavior of the NCAR AOGCM, does not seem to be realistic, the coefficients with $K_v = 0.5\ cm^2/s$ have been used as a lower limit. This range, while somewhat narrower than that given by AOGCMs, is still wider than that used by Wigley and Raper (1993), who arbitrarily chose coefficients twice as large and half as large as their standard value for upper and lower bounds or the uncertainty range. Since their UD/EB model has rate of heat uptake close to that of the MIT 2D model with $K_v = 5.0\ cm^2/s$, the corresponding range for the MIT 2D model would be from 2.5 to 10 cm^2/s. For climate sensitivity a range close to that suggested by the IPCC, namely 1.6 ˚C to 4.5 ˚C, has been used in this study. The simulations discussed below have been performed with a 1% per year increase in the CO_2 concentration, while all other forcings were held constant.

As can be seen from Figure 5.1, if the rate of heat uptake by the deep ocean is close to that matching the behavior of the NCAR model, the increase of the surface temperature will be significantly higher than the highest estimate of possible warming given by the IPCC. Actually, in the simulation with $\Delta T_{eq} = 4.5$ and $K_v = 0$ the surface temperature increase for years 91–100 of the integration, $\Delta T_{91\text{-}100}$, is 5.7 ˚C, compared to 4.6 ˚C for heat diffusion with $K_v = 0.5$ cm^2/s. One might argue that the combination of high sensitivity with low rate of heat uptake has quite a low probability. However, there is a noticeable difference in the simulations with $K_v = 2.5$ cm^2/s and $K_v = 12.5$ cm^2/s. For a climate sensitivity of 2.5 ˚C an increase in diffusion coefficient from 0.5 to 12.5 cm^2/s leads to a decrease in $\Delta T_{91\text{-}100}$ from 3.0 ˚C to 2.4 ˚C (not shown). This shows that for the upper part of the sensitivity range the impact of uncertainty in the rate of the heat uptake by the deep ocean on surface warming is comparable in magnitude with that of some other uncertainties, for example, uncertainty in radiative forcing associated with aerosols. In the case of low climate sensitivity the impact of the deep ocean on warming is much smaller (Figure 5.1).

Sea level rise due to thermal expansion (Figure 5.2), in contrast with the surface warming, is very sensitive to the rate of heat penetration into the deep ocean especially for low climate sensitivity. An increase in diffusion coefficient from 0.5 to 12.5 cm^2/s leads to a doubling of sea level rise by the end of the simulation with $\Delta T_{eq} = 1.6$ ˚C, while causing about a 40% change for $\Delta T_{eq} = 4.5$ ˚C. It is worth noting, that, while in the case of high climate sensitivity an increase in sea level due to thermal expansion may be somewhat offset (Wigley and Raper, 1993; IPCC, 1996) by the decrease in land ice melting (or *vice versa*), this would not be the case for low climate sensitivity.

6. Conclusions

The results presented from the simulations with the MIT 2D climate model show that it, while having some limitations compared to GCMs, reasonably reproduces the main features of the present-day climate, including seasonal variability. Both globally averaged values and zonal distributions of equilibrium changes in the different climate variables, such as temperature, precipitation, evaporation, and radiation balance at the surface, as produced by different versions of the 2D model in response to a doubling of the atmospheric CO_2, are similar to those obtained in simulations with different GCMs. Use of an artificial cloud feedback for changing the 2D model sensitivity does not produce any physically inconsistent results. The 2D model coupled with a diffusive ocean model, in spite of the simplicity of the latter, reasonably reproduces the transient behavior of different AOGCMs, at least for climate simulations on time scales of 100–150 years. These results together with the relatively moderate computer resource requirements make the 2D model a very useful tool for studying uncertainty in climate change both independently and as a part of the integrated framework of the MIT Joint Program on the Science and Policy of Global Change.

Our results show that there is a wide disagreement between coupled AOGCMs simulations on the rate of heat uptake by the ocean. The corresponding uncertainty in the surface warming is comparable in magnitude with the uncertainties in other parameters. The impact of oceanic heat uptake on the sea level rise is more complicated and strongly depends on chosen values of model parameters. As a whole, the impact of the uncertainty in oceanic heat uptake is significant enough to be taken into consideration in determining overall uncertainty in climate change.

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7. References

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