DESCRIPTION AND VALIDATION OF THE MIT VERSION OF THE GISS 2-D MODEL

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ABSTRACT

A significant number of long-term climate change simulations are to be carried out in the Integrated Framework of the MIT Joint Program on the Science and Policy of Global Change. Since Global Circulation Models (GCMs) require an enormous amount of computer time, the two-dimensional statistical-dynamic model developed by Stone and Yao has been chosen to be used for the initial stage of the Joint Program.

At MIT, the model has been modified to make it more suitable for the purposes of the Joint Program, including developing a new scheme for a surface flux calculation. A number of simulations with the modified version of the model have been performed in which a few schemes for cloud and ocean heat transport calculation have been tested. Comparisons of the results of the present climate simulations with observational data show that the model reasonably reproduces main features of zonally averaged atmospheric circulation. A climate sensitivity produced by the model coupled with a mixed layer ocean model in response to the doubling of the atmospheric CO_2 concentration lies in the range of the results obtained with GCMs. The results of the simulations with a gradual increase of the greenhouse gas concentrations in the atmosphere, in which diffusion of heat into the deep ocean was taken into account, are also similar to those obtained in the analogous simulations with GCMs. As a whole, presented results demonstrate that the modified version of the two-dimensional model can be successfully used for climate change predictions in the Integrated Framework of the Joint Program.

1. INTRODUCTION

The prediction of possible climate change caused by human activities is an important part of the MIT Joint Program on the Science and Policy of Global Change. The most sophisticated tools used for climate and climate change simulations are General Circulation Models (GCMs), which include both the atmosphere and ocean. A significant number of numerical simulations with GCMs have been carried out in recent years.

However, much more research has to be done before the model results can be considered robust. There is significant disagreement in the predictions of possible climate change obtained in simulations with different GCMs. For example, the change of the global averaged surface temperature in response to doubling of the CO_2 concentration in the atmosphere ranges from 1.9˚C to 5.2˚C (IPCC, 1990). Most of the variation in model sensitivity is caused by differences in the depiction of climate feedback processes (see, for example, Cess et al., 1990). This, in turn, is associated with the use of different parameterizations of physical processes (Cess et al., 1993). Since one of the main goals of the Joint Program is to evaluate uncertainties in climate change prediction (Jacoby and Prinn, 1994), simulations with different versions of an atmospheric model, covering the whole spectrum of climate sensitivity, have to be performed. A variety of scenarios for changes in greenhouse gas (GHG) concentrations also have to be considered. As a result, a significant number of climate simulations, each for 50 – 100 years, are to be carried out. This would be impossible with the use of GCMs, due to their enormous requirements of computer time, even on the most powerful super computers now available.

An alternative approach is to use simplified models. The two-dimensional (2-D) statistical-dynamical model developed at the Goddard Institute for Space Studies (GISS) is 23 times faster than the GISS GCM with the same latitudinal and vertical resolutions (Yao and Stone, 1987). Two-dimensional models certainly have some limitations compared to GCMs, such as an inability to simulate features of the atmospheric circulation caused by the temperature contrast between land and ocean, and the inability to take into account real topography. At the same time, preliminary studies performed with the 2-D GISS model suggest that after certain modification it could be used for climate change simulations.

At MIT a number of modifications have been made to the model to make it more suitable for the purposes of the Joint Program, and a comprehensive study of the model performance has been carried out. As a detailed description of the model has been published, only a brief one is given below, with emphasis on the changes made in the model at MIT. The modified MIT version of the model is hereafter referred to as the 2-D L-O model, where L-O refers to the inclusion of a real land-ocean distribution, which was not resolved in the GISS 2-D version. The results of the present climate simulations performed with the 2-D L-O model are compared below with observational data and the results of the analogous simulation with the GISS GCM. These comparisons show that the 2-D L-O model reasonably reproduces zonally averaged characteristics of the present climate.

Two kinds of climate change simulations carried out with the use of the 2-D L-O model are discussed below: an equilibrium climate change caused by doubling the atmospheric CO_2 concentration, and a transient change in response to a gradual increase of GHG content in the atmosphere. In the former, the atmospheric model has been coupled with a mixed layer ocean model; in the latter, diffusion of heat into the deep ocean has also been taken into account (see Hansen et al., 1988). The results of these simulations show that climate change, as far as globally averaged values and latitudinal distributions are concerned, is similar to that produced by GCMs. For example, the increase of surface temperature in the simulation with a doubled CO_2 concentration is 3.9˚C and lies in the range of results obtained with GCMs.

As a whole, the presented results demonstrate that the modified version of the 2-D L-O model can be successfully used for climate change simulations, at least for the initial stage of the Joint Program. At the same time, additional improvements are planned, such as the use of a more sophisticated ocean model. A fully three-dimensional (3-D) model may be used in future stages of the Joint Program, depending on available resources.

2. DESCRIPTION OF THE MODEL

As mentioned above, the original version of the 2-D model has been developed from the GISS GCM (Hansen et al., 1983). As a result, the model's numeric and parameterizations of physical processes, such as radiation, convection, etc., are closely parallel to those of the GISS GCM. The model solves the primitive equations as an initial value problem. The grid used in the model consist of 24 points in latitude, corresponding to a resolution of 7.826 degrees. The model has nine layers in vertical: two in the planetary boundary layer, five in the troposphere, and two in the stratosphere. The detailed description of the original version of the 2-D model is given in Yao and Stone (1987) and Stone and Yao (1987 and 1990). The important feature of the model, from the point of view of the Joint Program, 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, CFCs, etc., and twelve types of aerosols. The detailed description of radiation and parameterizations of other physical processes used in the model can be found in Hansen et al. (1983). At MIT a number of modifications have been made to the model, as described below.

2.1 Surface Fluxes

For the model's original purpose, it was possible to assume that the terrestrial boundary was all ocean. This is not an appropriate assumption given the use to which the model is put in the Integrated Framework of the Joint Program. The first necessary change was to include in the 2-D model a real land/ocean distribution (see Table 2.1). The modified land/ocean resolving 2-D L-O 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) as well as surface fluxes are calculated separately for each kind of surface. At the same time, the atmosphere is assumed to be well mixed horizontally; that is, air temperature, humidity, etc., are the same for the whole cell. The weighted averages of fluxes from different kinds of surfaces are used to calculate change of temperature, humidity, and wind speed in the first model's layer due to air/surface interaction. The same is true for the surface albedo used in radiative flux calculations. The descriptions of schemes used for ground temperature and moisture calculations are given in Hansen et al. (1983); calculation of sea surface temperature is described in Section 2.3.

Table 2.1 Fractions of latitudinal belt covered with land.

90N	82N	74N	67N	59N	51N	43N	35N	27N	20N	12N	4N
0.00	0.150	0.306	0.715	0.579	0.582	0.480	0.425	0.406	0.321	0.240	0.224
4S	**12S**	**20S**	**27S**	**35S**	**43S**	**51S**	**59S**	**67S**	**74S**	**82S**	**90S**
0.241	0.212	0.244	0.219	0.111	0.036	0.018	0.001	0.154	0.660	1.00	1.00

For reasons mentioned below, a new scheme for surface flux calculation has been developed by modifying the scheme used in the GISS GCM. The modified scheme, as well as the original, is based on the Monin-Obukhov similarity theory and uses the approximation for transfer coefficients derived by Deardorff (see Hansen et al., 1983). The fluxes of momentum, heat and moisture in the surface layer are calculated as follows:

$$\begin{aligned} \vec{\tau} &= -\rho\, C_n\, D_m \left|\vec{V}_s\right| \vec{V}_s \\ H &= c_p\, \rho\, C_n\, D_m \left|\vec{V}_s\right| \left(T_g - T_s\right) \\ E &= \beta\, \rho\, C_n\, D_m\, D_e \left|\vec{V}_s\right| \left(q_g - q_s\right), \end{aligned} \tag{2.1}$$

where $\vec{\tau}$ is surface stress, H is sensible heat flux, E is evaporation, ρ is air density, T is air temperature, q is specific humidity, $\vec{V}_s$ is surface wind, β is the ratio of available water to field capacity for top soil layer, and c_p is specific heat at constant pressure. Subscripts g and s correspond to ground and top of the surface layer. C_n, the transfer coefficient for neutral stratification, is

$$C_n = \frac{k^2}{\ln^2(z_o/z_s)} \ , \tag{2.2}$$

where the Karman constant, k = 0.35, z_o is the surface roughness, and z_s is the height of the surface layer; whereas, D_m, D_h, and D_e are functions of bulk Richardson number

$$Ri_s = \frac{z_s \, g \, (T_s - T_g)}{T_g V_s^2} \ , \tag{2.3}$$

namely:

$$\begin{aligned} D_m &= \left[\frac{(1 - a\,Ri_s)(1 - b\,Ri_s)}{1 - c\,Ri_s}\right]^{-1/2} \\ D_h = D_e &= 1.35\left[\frac{1 - d\,Ri_s}{1 - f\,Ri_s}\right]^{1/2} \ , \end{aligned} \tag{2.4}$$

for an unstably stratified surface layer ($Ri_s < 0$), and

$$\begin{aligned} D_m &= 1 + (11.2 + 90\,Ri_s)\,Ri_s^{-1} \\ D_h = D_e &= 1.35/(1 + 1.93\,Ri_s) \ , \end{aligned} \tag{2.5}$$

for stable stratification ($Ri_s > 0$). The coefficients in Equation (2.4) are functions of $\ln(z_o/z_s)$ and are given in Hansen et al. (1983). The height of the surface layer, z_s, is taken to be 10 m over ocean and sea-ice and 30 m over land. The surface roughness over land has been calculated from data on topography and vegetation (see Hansen et al., 1983 for details). It is assigned a value of 4.3×10^{-3} m for sea-ice and calculated by means of the Charnock formula for ocean:

$$z_o = 0.018 \times \tau/g \ , \tag{2.6}$$

where τ is the absolute value of surface stress from previous time step.

The main difference between the scheme used in the GISS GCM and this one is the assumption used to define values of variables on the top boundary of the surface layer. In the GISS GCM the surface layer is assumed to be in an equilibrium. The numerical realization of this assumption results in complicated algorithm including nested iterations. This algorithm, while used successfully in both the GISS GCM and 2-D model without land, produces computational problems when land is included. Because of this, the above mentioned assumption has been replaced by the assumption that the layer between the surface and the model's first level is well mixed. This assumption simply means that $\theta_s = \theta_1$ and $q_s = q_1$, where θ_1 and q_1 are potential temperature relative to the surface pressure and specific humidity on the model's first level.

The surface wind speed calculation also has been simplified compared to the procedure used in the GISS GCM. The absolute value of $|\vec{V}_s|$ is assumed to be equal to

$$|\vec{V}_s| = \sqrt{V_1^2 + \overline{(V_s'^2)}} \ , \tag{2.7}$$

where $\vec{V}_1$ is wind at the first model's level and $\overline{(V_s'^2)}$ is the zonal mean surface wind variance. The cross isobar angle, α, is a function of $|\vec{V}_s|$ and Ri_s, namely:

$$\alpha = \frac{0.0625 \times 2\,\pi}{1 + \left|\vec{V}_s\right| \alpha_o}\ , \tag{2.8}$$

if $Ri_s < 0$ and

$$\alpha = \frac{2\,\pi \left(0.09375 - \dfrac{0.03125}{1 + 4\,Ri_s}\right)}{1 + \left|\vec{V}_s\right| \alpha_o}\ , \tag{2.9}$$

if $Ri_s > 0$ (G. Russell, personal communication, 1993). Here α_o is an empirical coefficient that relates the cross isobar angle to the surface wind, which is taken equal to 0.3.

Finally, the components of surface wind are calculated as:

$$u_s = u_1 \cos\alpha - k_h\, v_1 \sin\alpha$$
$$v_s = v_1 \cos\alpha + u_1 \sin\alpha\ , \tag{2.10}$$

where $\vec{V}_1 = (u_1, v_1)$ is the wind on the first level, $k_h = 1$ in the Northern Hemisphere and -1 in the Southern Hemisphere.

2.2 Cloud Parameterization

Two types of clouds are taken into account in the model: convective clouds, associated with moist convection, and large-scale or supersaturated clouds, formed due to large-scale condensation. The amount of convective clouds in the given layer C_{mc} is proportional to mass flux due to moist convection through the lower boundary of this layer M:

$$C_{mc} = \gamma \cdot M\ , \tag{2.11}$$

where γ is assumed to be a function of latitude.

The amount of supersaturated clouds is a function of relative humidity,

$$C_{ss} = K \cdot \frac{h_{con} - h_c}{1 - h_c}\ , \tag{2.12}$$

where h_c is the critical value of relative humidity for cloud formation and h_{con} is the critical value of relative humidity for condensation. K and h_c are functions of height.

In the original version of the 2-D model, condensation occurs when relative humidity reaches 100%. As a result, the amount of precipitable water in the atmosphere obtained in the simulations with this version turns out to be larger than the observed value. At the same time, even in some GCMs with low horizontal resolution, condensation is allowed to occur in partly saturated areas, in order to take into account subgrid-scale variations of relative humidity. Such an approach seems to be even more appropriate in a zonally averaged model: moreover, a similar approach is used in the parameterization of moist convection (Yao and Stone, 1987). Therefore, the value of h_{con} = 90% has been chosen as the criterion for condensation. This small change has a very profound impact on the model's sensitivity, namely, if h_{con} = 100%, the model produces a negative cloud feedback; however, when h_{con} = 90%, the cloud feedback becomes positive.

Two versions of cloud parameterization have been used in the simulations described below. The first, developed by Yao (personal communication, 1993), is defined by the following values of the parameters: $\gamma = 7.50\ m^2\ skg^{-1}$ and

$$K\,(p) = \begin{Bmatrix} 0.60 \ \text{ for } \ p \le 400 \text{ mb} \\ 0.25 \ \text{ for } \ p > 400 \text{ mb} \end{Bmatrix} . \tag{2.13}$$

The value of γ used in this version is three times larger than the one used in GISS GCM. In the original version of this scheme h_c was equal to 0.85; however after the reduction of h_{con}, h_c was reduced to 0.8. In the second version:

$$\gamma\ (\varphi) = \begin{Bmatrix} 7.50\ m^2skg^{-1} \ \text{ for } \ |\varphi| \le 20^\circ \\ 3.75\ \ m^2skg^{-1} \ \text{ for } \ |\varphi| \ge 20^\circ \end{Bmatrix} , \tag{2.14}$$

K = 1, and h_c = 0.875. These two parameterizations, while producing a similar amount of total clouds, give quite different distributions of convective and supersaturated clouds.

2.3 Mixed Layer Ocean Model

In most of the simulations described below, the 2-D L-O model has been coupled with a mixed layer ocean model. In order toimulate the current climate, the equation for the mixed-layer temperature, T_o, must include a term, Q, 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:

$$C_w\, \rho_w \frac{\partial(T_o z_{ml})}{\partial t} - L_f \frac{\partial M_i}{\partial t} = f_o\, F_{ao}^{\downarrow} + f_i\, F_{io}^{\downarrow} + L_f \left(\frac{\partial M_i}{\partial t}\right)_{melt} + Q\ , \tag{2.15}$$

where C_w and ρ_w are specific heat capacity and density of salt water respectively, M_i is sea-ice mass, $z_{ml} = z_{ml}\ (t, \phi)$ is the mixed layer depth, L_f is the latent heat of freezing, t is time, $F_{ao}^{\downarrow}$ is the heat balance on the ocean surface, $F_{io}^{\downarrow}$ is the heat flux through the lower surface of sea-ice, f_o and f_i are fractions of grid cell covered by open ocean and sea-ice, and $(\partial M_i/\partial t)_{melt}$ is the rate of sea-ice melting. The Q-flux can be calculated from this equation using the results of the climate simulation with climatological sea surface temperature and sea-ice distribution. Namely:

$$Q = \left\{ C_w\, \rho_w \frac{\partial(T_o z_{ml})}{\partial t} - L_f \frac{\partial M_i}{\partial t} \right\}_{clim} - \left\{ f_o\, F_{ao}^{\downarrow} + f_i\, F_{io}^{\downarrow} + L_f \left(\frac{\partial M_i}{\partial t}\right)_{melt} \right\}_{fsst} , \tag{2.16}$$

where subscripts $_{clim}$ and $_{fsst}$ indicate observational values and values from simulation with fixed sea surface temperature (SST) and sea-ice. This procedure is essentially similar to the one used in the simulations described in Meleshko et al. (1991), except that in those simulations z_{ml} was constant, whereas here it is a function of time and latitude. The algorithm used for calculation of the thermal energy of the mixed layer with variable depth is described in Russell, Miller and Tsang (1985).

There is an essential difference between Q-flux used in a mixed layer ocean model and flux correction used in simulations with coupled atmosphere-ocean GCMs (see, for example, Gates et al., 1993). The latter is purely model error correction, while the former, in addition to such a correction, has, as mentioned above, certain physical meaning. In other words, if models were perfect, flux correction would be equal to zero, with Q-flux equal to observed ocean heat transport.

The implied annual ocean heat transport, calculated from the results averaged for ten years of the present climate simulation with fixed sea surface temperature and sea-ice distribution, is shown on Figure 2.1. This simulation has been carried out with the 2-D L-O model that incorporates the second cloud parameterization scheme (see next Section). The 2-D L-O model produces excessive heat transport in the Northern Hemisphere compared to most of the available observations. At the same time, the overall pattern of the ocean heat transport is in good agreement with observation (see Gleckler et al., 1994).

3. PRESENT CLIMATE SIMULATION

A significant number of present climate simulations have been performed with the 2-D L-O model, and different versions of the above described parameterization schemes have been tested. Most of the parameterizations were tested in the simulations with fixed SST and sea-ice. However, our goal was to develop a model capable not only of simulating reasonably well the present climate but also of reproducing the climate change pattern obtained in simulations with GCMs. Because of this, some changes in the parameterization were made on the basis of the analysis of the model's response to the doubling of the CO_2 concentration. As a result, several versions of the model with different sensitivities have been developed. The results of two present climate simulations, carried out with versions of the 2-D L-O model incorporating different cloud parameterization schemes, are described in this Section.

In both simulations the 2-D L-O model coupled with the mixed layer ocean model has been integrated for fifty years. The results averaged for the last ten years are shown below. The Q-fluxes have been calculated from simulations with fixed sea surface temperature and sea-ice performed with the corresponding version of the 2-D L-O model, as described in the previous Section. As the 2-D model tends to produce a two-grid-size nose (see Yao and Stone, 1987), the Q-fluxes have been smoothed with a three-point filter.

Since there is essentially no difference in the climates simulated with the different versions of the model, climate characteristics are shown for the simulation with the second cloud parameterization scheme only, unless otherwise indicated.

3.1 Temperature

A cold bias in the air temperature is a common feature of almost all existing GCMs, despite differences in horizontal and vertical resolutions, physical parameterizations, etc. (Boer et al., 1992; Hurrell, 1995; Murphy, 1995). Differences between simulated and observed (Oort, 1983) temperature for Northern Hemisphere winter (DJF) and summer (JJA) show that the 2-D L-O model shares this feature (Figures 3.1 and 3.2) as well. The most significant errors in the simulated temperature are seen in the stratosphere in the polar regions (especially in a summer hemisphere). At the same time, the tropospheric temperature is well simulated by the model, except near the winter pole.

3.2 Zonal Wind

The 2-D L-O model reproduces reasonably well the main features of the zonal wind distribution, such as the magnitudes and locations of jet streams and the easterlies in low latitudes

(Figures 3.3 and 3.4). The fact that the winter jet in the Northern Hemisphere is not closed is a result of the above mentioned deficiencies in the simulated air temperature, namely, too weak of an inverse latitudinal temperature gradient in the stratosphere. A reasonable simulation of the surface wind is of particular importance for possible coupling with a dynamic ocean model, as it strongly affects atmosphere ocean interaction. The simulated zonal surface stress is shown in Figure 3.5 together with the observed values of Hellerman and Rosenstein (1983). Most of the existing atmosphere/ocean GCMs (AOGCMs), except the United Kingdom Meteorological Office (UKMO)AOGCM, tend to underestimate, to different extents, the surface stresses over the oceans (Gates et al., 1993; Murphy, 1995). The 2-D L-O model also underestimates surface stress over open ocean (Figure 3.5b). At the same time it reproduces well the global stress associated with westerlies in mid latitudes and even overestimates easterly stress in low latitudes (Figure 3.5a). This disagreement arises in part because the same value of surface wind is used for the whole latitude circle, while, according to observations, the surface wind is usually stronger over ocean than over land. As the surface roughness of land is significantly larger than that of ocean, it leads to larger stresses over land. The dependence of the ocean surface roughness on the surface wind amplifies this discrepancy.

3.3 Stream Function and Eddy Transport

Comparison of the calculated meridional stream function (Figures 3.6 and 3.7) with observations shows that the simulated Hadley circulation is much weaker than observed during the Northern Hemisphere winter, while being about right in summer. The location of the rising branch is also closer to observation in Northern Hemisphere summer.

Eddy fluxes of momentum, heat and moisture (not shown) are very similar to those produced by the original GISS version of the 2-D model. A detailed discussion and validation of these can be found in Stone and Yao (1987 and 1990).

3.4 Clouds and Radiation

The zonal distribution of annual mean super saturated and convective clouds obtained in the two above mentioned simulations are shown in Figure 3.8. Hereafter, the acronym S1 is used for the simulation with the second cloud parameterization scheme, whereas S2 indicates the first. The results of the present climate simulation with the GISS Model II (Hansen et al., 1983) are also shown for the comparison. In spite of the significant differences in super saturated and convective clouds, the total cloud cover is similar in both simulations (Figures 3.9 and 3.10). Both versions of the model significantly underestimate cloud cover in the vicinity of 20˚S during Northern Hemisphere winter and near 30˚N during Northern Hemisphere summer. (A list of sources of observational data is given in Section 6.) This is in part because of the excessive zonality of the circulation in a 2-D model (cf. the GCM results in the same figure). This deficiency also reveals itself in the simulated planetary albedo and net radiation at the top of the atmosphere (Figures 3.9 through 3.12), especially during Northern Hemisphere summer. At the same time, annual mean net radiation at the top of the atmosphere and, as a result, implied northward energy transport are in good agreement with observation (Figure3.13). The data on energy transport may be found in Gleckler et al. (1994).

Another variable that characterizes cloud impact on the radiation is cloud radiative forcing, that is, the difference between total radiation fluxes and those for clear sky condition. Radiation fluxes for clear sky have been calculated by Method II, as defined by Cess and Potter (1987). It is worth noting that both versions of the model underestimate outgoing longwave radiation for clear sky condition because of the cold bias (see Section 3.1). As one can see the deviation of the simulated cloud radiative forcing from observed values is not well correlated with that for the total cloud cover (Figures 3.14 and 3.15 versus Figures 3.9 and 3.10). For example, low absolute values of cloud radiative forcing in the area from 20˚N to 40˚N in JJA (Figure 3.15) are associated with the underestimated cloud cover in this region (Figure 3.10). However, the simulated cloud radiative forcing exceeds the observed forcing in low latitudes of the Southern Hemisphere despite the deficit of total clouds. The latter, together with slightly lower than observed values of longwave cloud radiation forcing, indicates a relatively excessive amount of low clouds. There are no observational data on vertical distribution of cloud, but comparison with GCM results confirms this (see, for example, Senior and Mitchell, 1993). However, it is necessary to keep in mind that the cloud impact on radiation fluxes is different in 2-D and 3-D models because of different overlapping.

As mentioned, cloud feedback is one of the main reasons for differences in model depictions of climate response to increase in GHG concentrations. Seasonal differences can give some information on this matter. The 2-D L-O model overestimates the seasonal change in total cloud cover in subtropical regions of both hemispheres (Figure 3.18). Neither the 2-D L-O model nor the GISS GCM reproduces the seasonal cloud change in the tropics associated with the shift of the Intertropical Convergence Zone. Nevertheless, the overall pattern of seasonal cloud changes simulated by the 2-D L-O model, especially in the simulation S1, is quite similar to the observed one. The same is true for the seasonal change in cloud radiative forcing.

3.5 Hydrological Cycle

The 2-D L-O model's ability to simulate the hydrological cycle reasonably is important both for coupling with an ocean model and for predicting a possible climate change and its impact on the ecosystem. The zonally averaged precipitation and evaporation simulated by the 2-D L-O model are shown in Figures 3.18 and 3.19. The northward shift of the maxima of precipitation in DJF compared to observation is consistent with the corresponding shift of the location of the rising branch of the mean meridional circulation (Figure 3.6). Both the 2-D L-O model and the GISS GCM have difficulty matching the observed precipitation, particularly in DJF. As one can expect, the agreement between the 2-D L-O model simulation and observed precipitation in the equatorial region is much better in JJA. The 2-D L-O model significantly overestimates precipitation in the areas of the descending branch of mean meridional circulation in both seasons; however, there is noticeable disagreement between observational data from different sources as well. To illustrate this the results of the simulation S1 are shown in Figure 3.20 together with the data of Leemans and Cramer (1990) and Shea (1986). The overall pattern of zonally averaged evaporation is reasonably well reproduced by the model.

One of the driving forces of an ocean model is the water flux from the atmosphere, that is, precipitation minus evaporation. In comparing the model's results with observations (Figure 3.21) one has to keep in mind that most of the above mentioned discrepancy in observations is because of poor data over ocean. In addition, discrepancies in data on precipitation minus evaporation is much larger than in precipitation and evaporation separately.

3.6 Sea Surface Temperature and Sea-Ice

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, essentially associated with longitudinal variability, can not be simulated by a 2-D model. Because of this, data used in the simulations with prescribed SST and sea-ice have been adjusted. Namely, if less than 10% of the ocean surface is covered by ice, and SST is above 0˚C, ice is removed. If ice covers more than 10% of the ocean, then surface SST is set to 0˚C if mass of ice is decreasing, and to –1.56˚C otherwise. The adjusted data are shown in Figures 3.22 and 3.23 instead of direct observations.

As one can see, despite the use of a Q-flux in the mixed layer model (Equation 2.15) there are some differences in prescribed and simulated values of SST and sea-ice cover. In particular, the 2-D L-O model overestimates sea-ice extent in the Southern Hemisphere in both seasons, while slightly underestimating it in the Northern Hemisphere in JJA. There are two main sources for these differences. First, SST and sea-ice mass are calculated every hour, whereas the Q-flux is calculated from monthly mean data, and the values used in Equation 2.15 are obtained by interpolation. The second, and perhaps more important, reason is that it is not sea-ice mass that is needed in the atmospheric model, but sea-ice area and depth. Both are prescribed in the simulation with prescribed SST and sea-ice. On the other hand, the mixed layer ocean model predicts sea-ice mass, which is then used to calculate ice area and depth. So, even if sea-ice mass were simulated perfectly there could still be disagreement between the prescribed and simulated ice cover and ice depth at some grid points. The latter, in turn, will cause differences in, for example, heat fluxes at the atmosphere ocean interface and horizontal heat transport. It should be mentioned that the depth of the simulated sea-ice is less than what was used in the run with prescribed SST and sea-ice, except at the North Pole.

As a whole, the above given comparison of the model’s results with the observational data shows that the 2-D L-O model reasonably well reproduces the major features of the present climate state. There are some essential 3-D features of the atmospheric circulation, such as Walker circulation, that, of course, can not be simulated by a 2-D model. However the depiction of the zonal averaged circulation by the 2-D L-O model is not very different from that by GCMs. Since the model is to be used for climate change prediction, it is noteworthy that the seasonal climate variation is also reproduced quite well.

4. RESPONSE TO CHANGES IN CO_2 AND OTHER GHG CONCENTRATIONS

In this Section attention is paid mainly to the equilibrium climate change, that is, the model’s response to a doubling of atmospheric CO_2 concentration. A significant number of such simulations have been performed with GCMs (IPCC, 1990). Despite the differences in model results, simulations of this kind are useful tests for model validation.

The results of simulations with two versions of the 2-D L-O model are shown below. In both cases the model has been integrated for 50 years after doubling of the CO_2 concentration. The difference between the model data averaged over the last decade of the perturbed simulation, and the last decade of the corresponding control run, represents the simulated equilibrium climate change. As mentioned in the Section 1, the equilibrium change of the surface air temperature produced by different GCMs ranges from 1.9˚C to 5.2˚C (4.2˚C for the GISS GCM), the corresponding values obtained in simulations S1 and S2 are 3.93˚C and 3.68˚C, respectively. In this Section, S1 and S2 denote climate change simulations performed with the corresponding version of the 2-D L-O model, rather than present climate simulations. The change in precipitation caused by a doubling of CO_2 concentration is shown on Figure 4.1 as a function of surface warming. Data for simulations with different GCMs are taken from Washington and Meehl (1993). The increase in precipitation varies from 3 to 15%, being equal to 11.7 and 10.9% in the simulations with the 2-D L-O model. In Table 4.1 changes in the components of the surface energy budget as obtained in simulations with four GCMs (Boer, 1992) are shown together with the results from simulation S1 with the 2-D L-O model.

Table 4.1 Change in globally and annually averaged terms of the surface energy budget due to a doubling of the CO_2 concentration.

	NCAR	GFDL	GISS	CCC	MIT 2-D L-O
R	5.5	6.4	7.0	1.4	7.8
LE	– 7.1	– 7.1	– 10.4	– 3.0	– 9.3
H	2.7	0.6	3.3	1.6	1.4
N	1.0	– 0.1	– 0.1	0.0	– 0.1
S	1.3	1.5	2.9	– 2.4	3.1
F	4.2	4.9	4.1	3.8	4.7
B	– 0.036	– 0.017	– 0.061	– 0.029	– 0.047

Here R = S + F is the net radiation at the surface; S and F are the short and longwave radiation components respectively; LE and H are the turbulent fluxes of latent and sensible heat; N is the net energy flux; and B = LE ⁄ H is the Bowen ratio. As can be seen in the table, changes of the globally averaged values of different climate characteristics fall into the range produced by GCMs, and most are close to those obtained in the simulation with the GISS GCM (Hansen et al., 1984). The latter is not surprising given the origin of the 2-D L-O model. However, latitudinal distributions of the change simulated by the 2-D L-O model and GISS GCM do differ. For example, the increase of air surface temperature (Figure 4.2) produced by the GISS GCM varies with latitude less than in the 2-D L-O model. The GISS GCM produces larger warming and, as a result, larger increase in precipitation in the tropics than the 2-D L-O model, whereas, the opposite is true for surface air temperature increase in high latitudes, especially in the Northern Hemisphere. The latitude-height distribution of the CO_2-induced change of zonal mean air temperature calculated in simulation S1 (Figure 4.3) resembles more that produced by the GFDL GCM (Wetherald and Manabe, 1988) than by the GISS GCM (Hansen et al., 1984). The difference in the latitudinal distribution of temperature changes simulated by the GISS GCM and the 2-D L-O model is because of different strengths of the sea-ice and snow feedback (Figure 4.4). A detailed climate feedback analysis showed that the surface albedo feedback produced by the 2-D L-O model, associated mainly with changes of snow and sea-ice cover, is twice as large as that for the GISS GCM (Sokolov and Stone, 1994). The excessively strong sea-ice feedback is caused, in part, by the deficiency in the sea-ice simulation mentioned in previous Sections. A study of possible improvements in sea-ice simulation is in progress. Differences in the sea-ice and snow feedback also partly account for the different sensitivity of the two versions of the 2-D L-O model.

The CO_2-induced changes of clouds, planetary albedo, and radiation fluxes at the top of the atmosphere are shown on Figures 4.5 and 4.6. These changes bear a general resemblance to the results of the GISS GCM. However, the cloud feedback in the 2-D L-O model is weaker than in both the GISS and GFDL GCMs, even though the cloud cover changes are larger. The ratio of surface temperature change in a simulation with interactive clouds to one with fixed clouds is 1.75 and 1.25 for the GISS and the GFDL GCMs respectively, and 1.1 for both versions of the 2-D L-O model. The relatively small cloud feedback in the 2-D L-O model is a result of compensation between the changes in shortwave and longwave radiation caused by CO_2-induced changes of clouds. It is worth recalling that there is significant disagreement between GCMs in the strength, and even sign, of cloud feedback (Cess et al., 1990, Senior and Mitchell, 1993).

In spite of differences in feedbacks between the 2-D L-O model and some GCMs, the above comparison shows that the 2-D L-O model is capable of reproducing the behavior of a GCM as far as changes of zonal mean and global values of climate variables are concerned. A number of versions of the 2-D L-O model with different sensitivities have been developed by artificially changing the cloud feedback. For example, a version like S1, but with fixed supersaturation clouds, produces surface warming of 5.2˚C, which is the upper limit of sensitivities obtained in simulations with GCMs.

The results of the simulations with the present version of the 2-D L-O model have been used in a number of impact studies performed in the Integrated Framework of the Joint Program. Therefore, it seemed worthwhile to give the above comparison. However, it is necessary to keep in mind that studies on further improvement of the 2-D L-O model are underway and the version of the model finally chosen as a standard one for the Joint Program may be somewhat different from that described in this paper.

Simulations with different GCMs give different rates of global warming caused by a gradual increase of GHG concentrations (Murphy and Mitchell, 1995; IPCC, 1992). As shown by Hansen et al. (1984), a delay in warming depends on both the model's sensitivity and the rate of heat uptake by the deep ocean. A number of simulations with different versions of the 2-D L-O model are being carried out in an attempt to match the behavior of different GCMs. Coefficients of heat diffusion into the deep ocean in the standard version of the 2-D L-O model are based on the values used in the GISS GCM, which is based on measurements of the mixing of tritium into the deep oceans (Hansen et al., 1984). They vary with latitude from 0.3 – 0.5 cm^2s^{-1} in the equatorial region to 5 – 9 cm^2s^{-1} in high latitudes, with the global averaged value equal to 2.5 cm^2s^{-1}. As shown in Figure 4.7, the response of a version of the 2-D L-O model with these values of diffusion coefficients and a sensitivity of 3.77˚C to 1% per year increase in the CO_2 concentration is similar to that obtained in the same scenario with the GFDL GCM, which has a similar sensitivity (Manabe et al., 1991). This comparison illustrates that the transient behavior of GCMs can be reproduced by choosing an appropriate heat diffusion. This is important for evaluating uncertainties in transient climate change predictions with the 2-D L-O model.

5. SUMMARY

The results of simulations with the modified 2-D L-O model show that it, while having some limitations, reasonably well reproduces the main features of the present climate. The model's response to both an instant and a gradual increase in the concentration of greenhouse gases is consistent with the results of the simulations with GCMs. This shows that the model can be used for state-of-the-art climate change predictions in the Integrated Framework of the Joint Program. These results, together with the moderate computer resources requirement, makes the 2-D L-O model a very useful tool for studying uncertainties in global change. At the same time, a number of improvements can be made to the model, and studies of some are underway.

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7. LIST OF DATA SOURCES

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Earth Radiation Budget Experiment (ERBE)

Longwave and shortwave radiation fluxes for clear sky condition (Figures 3.14-3.16)
Earth Radiation Budget Experiment (ERBE)