CHAPTER II

A BRIEF DESCRIPTION OF THE UIUC 24–LAYER ST–GCM

This chapter introduces the UIUC 24–layer stratosphere–troposphere general circulation model (ST–GCM) used in this study. A detailed description of the model development and validation is presented in Appendix A.

A. History of the UIUC GCMs

The 24–layer ST–GCM is a descendent of the 2–layer atmospheric GCM developed in the late 1960's and early 1970's by Arakawa and Mintz at UCLA (Gates *et al.* 1971; Ghan *et al.* 1982) and subsequently developed and used by Schlesinger. The 2–layer AGCM has been used for many simulation studies, including the onset of the last ice age (Schlesinger and Verbitsky 1996) and the equilibrium climate change induced by doubling the CO_2 concentration (Schlesinger and Zhao 1989). Beginning in 1984, Schlesinger and Oh developed a 7–layer version of the AGCM, which was used by Oh for his Ph.D. research at Oregon State University to develop and test a physically based parameterization of clouds and their radiative interactions (Oh 1989). The 7–layer AGCM, with its top at 200 hPa, differs from the 2–layer AGCM mainly in its vertical resolution and the treatment of radiation, clouds, precipitation and the planetary boundary layer. Wang (1996) developed an 11–layer AGCM by extending the model top of the 7–layer AGCM to 50 hPa and adding a few layers in the lower stratosphere. The 11–layer AGCM possesses the same dynamic and physical features as the 7–layer AGCM, but is significantly improved in simulating the present climate, especially the tropical intraseasonal oscillation (Wang and Schlesinger 1999).

The 24–layer ST–GCM has been under development since 1994 primarily based on the 11–layer AGCM. A main purpose of developing such a model was to calculate the radiative forcing and simulate the climate changes induced by the Pinatubo eruption.

B. Basic Structure and Newly Implemented Components

The model has a horizontal resolution of 4°–latitude by 5°–longitude (Appendix A) and uses staggered B–grid for finite differences (Arakawa and Lamb 1977). Vertically the model extends from the earth's surface to 1 hPa (Appendix A) and uses sigma (σ) as its vertical coordinate. Given a surface pressure of 1000 hPa, there are 10 layers above 100 hPa with constant log–pressure thickness and 14 layers below 100 hPa with prescribed pressure values.

During the development of the 24–layer ST–GCM, new parameterizations were adopted for the transfer of both solar and longwave (terrestrial) radiation. The old parameterizations used in the UIUC 7–layer and 11–layer AGCMs proved to be unsatisfactory for the 24–layer ST–GCM in the stratosphere because of inaccurate heating rates and cooling rates. The inaccuracy of the longwave cooling resulted mainly from the absence of the Doppler broadening of the absorption lines of water vapor, carbon dioxide and ozone. A new parameterization for longwave radiation, originally developed by Chou and Suarez (1994), was adopted and modified for use in the UIUC 24–layer ST–GCM. It computes absorption and emission of terrestrial radiation due to water vapor, carbon dioxide, ozone, clouds, aerosol, and the trace gases N_2O, CH_4, CFC–11, CFC–12, and HCFC–22. It contains 9 broad bands ranging in wavenumber from 3000 cm^{-1} to infinity. This parameterization is capable of computing the cooling rate in clear sky accurately for both the middle and lower atmospheres (from 0.01 hPa to the surface) with errors less than 0.4°C/day (Chou and Suarez 1994). A new parameterization for solar radiation, originally developed by Chou and Suarez (1999), was also adopted and modified for use in the UIUC 24–layer ST–GCM. It computes the absorption by water vapor, ozone, carbon dioxide, oxygen, clouds and aerosols, and the scattering by clouds, aerosols and molecules (Rayleigh scattering). There are 8 bands in the ultraviolet and visible spectral regions (0.175–0.7 µm) and three bands in the near–infrared and thermal–infrared regions (0.7–10.0 µm).

The new and old parameterization schemes were compared by running 1–D radiative–transfer models under different clear–sky atmospheric conditions (Appendix A). For longwave radiation, for the cases of mid–latitude summer and tropics, the old scheme overestimates the cooling rates in the stratosphere with the largest error of about 2°C/day occurring in the middle stratosphere near 10 hPa, and underestimates the cooling rates in the troposphere, except near the surface, with the largest error of about 1.2°C/day occurring near 300 hPa. For the case of sub–arctic winter, the old scheme slightly underestimates the cooling rate in the middle troposphere, but severely overestimates the cooling rate in the upper stratosphere, with the largest error reaching 4.8°C/day. For solar radiation, the new and old schemes result in nearly identical heating rates in the troposphere. In the stratosphere, the heating rate computed by the old scheme is generally smaller than that computed by the new scheme and the discrepancy increases with height. At the model top the discrepancies are 1.1°C/day for the case of mid–latitude summer, 1.5°C/day for the case of tropics and 0.4 °C/day for the case of sub–arctic summer.

To study the climatic impacts of anthropogenic aerosols in the troposphere and volcanic aerosols in the stratosphere, a routine was developed to incorporate the radiative effects of these aerosols into the 24–layer ST–GCM. Both the scattering and absorption by aerosols are included in the solar radiation parameterization, while absorption and emission are included in the terrestrial radiation parameterization. For anthropogenic aerosols, radiative properties (specific extinction, single–scattering albedo and asymmetry factor) were calculated off–line by a Mie scattering model (see Appendix B), which was developed for an intercomparison project for studying the direct shortwave radiative forcing by sulfate aerosols (Boucher *et al.* 1998). The indirect radiative effect of anthropogenic aerosols in the troposphere is also parameterized by empirically relating the cloud–droplet number concentration to sulfate aerosol mass concentration (Boucher and Lohmann 1995). This aerosol radiation package has also been installed in the UIUC 11–layer GCM to study the global and regional climate changes induced by the direct solar radiative forcing of anthropogenic sulfate aerosol (Schlesinger *et al.* 1997a).

To be consistent with the new radiation schemes, cloud–radiation interaction was also modified. Slingo (1989)'s scheme on the shortwave radiative properties of liquid–water clouds, which depend on liquid–water path and the equivalent radius of the drop–size distribution (r_e), was adopted. The drop–size distribution (r_e) is determined by the in–cloud liquid–water content and cloud–droplet number concentration. Shortwave radiative properties of ice clouds are also functions of the ice–water path and ice crystal effective size (Chou *et al.* 1996). The effects of clouds on terrestrial radiation are included by introducing a mean flux transmittance, which is the multiplication of the gaseous transmittances and a cloud–related coefficient. This coefficient is calculated for each GCM layer and conveys information about cloudiness, cloud optical thickness, and cloud overlapping (Chou and Suarez 1994). Clouds are grouped into three categories – high clouds above the 16th σ–layer of the model, middle clouds between the 16th and 19th σ–layers, and low clouds below the 19th σ–layer. Clouds within each category are assumed to be maximally overlapped, and clouds among different categories are assumed to be randomly overlapped.

Processes such as stably stratified airflow over irregular terrain, moist convection, the development of dynamical instabilities of the Kelvin–Helmholtz type, geostrophic adjustment and frontal zones can produce gravity waves and transfer mean momentum between the troposphere and the stratosphere/mesosphere. The drag effect of gravity waves with spatial scales smaller than those resolved by an AGCM's grid is crucial for the model's performance (Hamilton 1996). We included Palmer *et al.* (1986)'s parameterization of orographically excited gravity–wave drag (GWD) in the 24–layer ST–GCM. Sensitivity studies were carried out to modify this parameterization so that it can best improve the model's performance. The inclusion of the parameterization in the model improves significantly the simulated tropospheric sub–tropical jets and sea–level–pressure centers in both the Northern and Southern Hemispheres; however, it over weakens the northern–hemisphere polar–night jet. It should be pointed out that how to parameterize the tropospheric/lower stratospheric GWD in GCMs is still not settled. It is even less clear than how to parameterize the mesospheric GWD. Some improved GWD parameterizations considering the spectral property and different sources of subgrid–scale gravity waves have been developed recently (e.g., Alexander and Dunkerton 1998; Hines 1997a; b). These parameterizations should be tested in the 24–layer ST–GCM in the future to further improve its simulation in the stratosphere.

Many parameters in the model have been tuned to enable it to best simulate the observed present climate. These parameters include, for instance, the thresholds of relative humidity for large–scale and convective precipitation, the characteristic time for the evaporation of precipitation, and the autoconversion rates of cloud water to precipitation for cumuloform and stratiform clouds.

C. Performance in Simulating the Present Climate

To validate the model's performance, a 15–year control simulation has been conducted with prescribed climatological sea–surface temperature and sea–ice distributions (Gleckler 1999). The model simulates well the geographical distributions and their seasonal variations of surface–air temperature and precipitation. The simulated large precipitation in the tropics moves along with the ITCZ. The simulated cloud cover matches the observed generally well except in the southern high latitudes, where the simulated is about 20% to 30% less than observed. The model is capable of reproducing the observed geographical distribution and seasonal variation of longwave cloud–radiative forcing (LW CRF), especially in the tropics. It is found that the LW CRF in the tropics largely depends on the emitting temperature of convective clouds at cloud top, which, in turn, depends on a threshold relative humidity that controls the onset of penetrating convection in the model (Wang and Schlesinger 1999). The model captures the basic geographical distribution and seasonal variation of the shortwave CRF. Large biases occur near 60°S in January because the simulated cloud cover has large errors there. Over the warm pool, the simulated annual–mean absorption of solar radiation by clouds is 35 W/m^2 smaller than observed.

The model simulates correctly the location of the tropical tropopause, the tropospheric mid–latitude temperature gradients and the sub–tropical jet streams. The model also simulates well the reversal of the observed pole–to–pole temperature gradient between summer and winter in the stratosphere and the southern-hemisphere polar–night jet. The model captures the location and phase of the semi–annual oscillation near the stratopause. Stratospheric sudden warmings are also detected in the northern–hemisphere middle stratosphere. However, a number of deficiencies exist. The model has a colder–than–observed lower polar stratosphere but a warmer–than–observed middle and upper stratosphere in the polar–night regions, especially in the Northern Hemisphere. The simulated northern–hemisphere polar–night jet is too weak compared to the observations, and the jet core is shifted towards the equator. These warm biases and the weaker–than–observed polar vortices have not been found in most other GCMs, which usually show systematic cold biases and over–intensified polar vortices. Sensitivity studies indicate that they are related to the use of the GWD parameterization in the 24–layer ST–GCM. Analyses of the residual circulation indicate that the model simulates reasonably well the two–cell Brewer–Doboson circulation in the stratosphere and its reversal between the two solstice seasons. The easterly forcing in the middle to upper stratosphere in the winter hemispheres generated by the model–resolved eddies is comparable in magnitude to those simulated by a few mesospheric GCMs. However, the forcing in the lower stratosphere seems to be too strong.

In summary, the UIUC 24–layer ST–GCM has significantly improved the simulation in the troposphere and near the surface with newly implemented parameterizations and modifications compared to its ancestors. The model has been coupled with the UIUC Atmospheric Chemical Transport Model (ACTM) in an off–line mode to simulate the distributions of source gases and ozone in the stratosphere (Rozanov *et al.* 1999a; b). It has also been used to reconstruct the radiative forcing of historical volcanic eruptions (Andronova *et al.* 1999). A 17–year transient simulation from 1979 to 1995 has also been performed using the model for the Second Atmospheric Model Intercomparison Project (AMIP–II) (Gleckler 1999). Based on the detailed validation in Appendix A and its applications in the aforementioned experiments, we may conclude that the UIUC 24–layer ST–GCM is suitable for the study of the radiative forcing and climatic impact of the Pinatubo eruption.