Simulation of the Martian Atmosphere Using a CCSR/NIES AGCM

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(Manuscript received 29 March 2002, in final form 3 September 2004)

Abstract

The new simulation of Martian general circulation based on CCSR/NIES AGCM has been performed and the numerical results analyzed. We attempt to reproduce the atmospheric states of Mars which have been observed by the Mars Global Surveyor, and Viking by introducing three kinds of time- and latitude-dependent dust opacity scenarios which are made to be consistent with the past observational results, as the observed dust distribution during spring and summer in south hemisphere varies largely at each year and so does observed atmospheric conditions. The model with TES2 dust scenario, which is the dust distribution based on the observation by the Mars Global Surveyor in 1999, generally reproduces distributions of temperature and zonal wind observed by the Mars Global Surveyor in 1999, though the temperature tends to be lower at high altitude. The maximum value of CO₂ ice thickness at polar regions relatively reproduces the observation, but the edge of the northern and southern seasonal caps are outside of the observational results during spring, which results in lower surface pressure especially in northern spring. Annual variations of the amplitudes of the diurnal and semidiurnal tide are qualitatively comparable between our model with VIK1 dust scenario, which is the dust distribution based on the observation by Viking Orbiter in 1977, and observation by Viking Lander 1 in 1977. The baroclinic waves with consistent periods, wavenumbers and phase speeds with observational results by Viking Lander 2 in 1977 are reproduced in the autumn, though in the winter, during the second dust storm, and in the spring, the phase speeds are faster than observation.

1. Introduction

Atmospheric conditions on Mars are quite different from those of the Earth. The dominant atmospheric component is CO₂ (95.3% of the atmosphere) and there is little water. The averaged surface pressure on Mars is about 6.1 hPa, and surface pressure varies annually with the range of 25% of the averaged pressure, because the atmospheric component of CO₂ condenses at polar regions (e.g. James et al. 1992).

Dust in the atmosphere plays very important roles in Martian meteorology, as well as radiative effects of CO₂. Atmospheric temperature and structure relate closely to dust opacity in

the atmosphere. Especially in southern spring and summer the dust storm activity becomes strong in every Martian year, because Mars' eccentricity (0.0934) is large and perihelion of Mars is near the southern summer solstice (aerocentric longitude $L_s = 250^\circ$ at perihelion, where $L_s = 90^\circ$ and $L_s = 270^\circ$ correspond to northern and southern summer solstices, respectively). The amount of insolation at southern spring and summer is about 1.3 times larger, so the atmospheric convection becomes stronger than that in northern spring and summer. The scale of the dust storm which occurs in this period differs in each year (from regional to planet-encircling), and hence atmospheric temperature has strong year-to-year variability according to the dust opacity. In contrast, in northern spring and summer the dust opacity and air temperature are lower and exhibit high year-to-year repeatability. Ob-

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served zonal-mean infrared ($9\ \mu\text{m}$) dust opacity at the equator, using a modified version of the retrieval scheme by Martin (1986), is around 0.05 in northern spring and summer at each year observed by Mariner 9, Viking Orbiters and Mars Global Surveyor (MGS). In southern spring and summer the observed maximum dust opacity at each year varies between 0.25 and 1.4 at the equator, according to the scales of occurred dust storms (Liu et al. 2003).

In the northern and southern hemispheres, CO_2 begins to condense out of the atmosphere at high latitudes in autumn as the regions encounter increasing darkness, depositing a thin veneer of dry ice, and the surface frost zone expands to mid-latitudes in late winter. In early spring, as the surface receives more sunlight, surficial CO_2 begins to sublime and the frost zone retreats to high latitudes. The variation of elevation due to CO_2 polar cap observed by the Mars Orbiter Laser Altimeter (MOLA) on the MGS is 1.0–1.5 m in both poles within a year (Smith et al. 2001b). In north polar regions the seasonal cap regresses with remaining relatively circular and uniform shape during northern late winter to spring. The recession of cap edge latitude versus L_s during this time is roughly represented as a linear function, though there is some slight interannual variability, by the observations of the Hubble Space Telescope (HST) and the Mars Orbiter Camera (MOC) on the MGS (Cantor et al. 1998; James and Cantor 2001; Kieffer and Titus 2001). In contrast, the south polar cap regresses axisymmetrically, especially after $L_s = 240^\circ$ (e.g. Kieffer et al. 2000).

The existence of large thermal tides plays an important role in Martian meteorology. Since the convective thermal drive on Mars is about ten times as effective as the diurnal forcing of Mars (per unit mass), because the atmosphere is very thin and has a very small heat capacity, the tidal response may be expected to scale up accordingly. In addition, the Martian atmosphere is strongly heated internally because of suspended dust, leading to a further enhancement of the tides, the enhancement varying with the dust load (Leovy 1979). In Viking observations, the amplitudes of diurnal and semi-diurnal components of the surface pressure variation were around 1 percent of the mean pressure at Lander 1 (VL1, 22.5°N , 48°W) and

0.5 percent at Lander 2 (VL2, 48°N , 226°W) when it is without global dust storms, and they became 3–4 times larger during the global dust storms (Leovy and Zurek 1979; Tillman 1988). Also about wind speeds, it was observed by VL2 that diurnal and semidiurnal variations became large during global dust storms (Leovy and Zurek 1979; Leovy 1981).

In northern midlatitude winter, the oscillations with the period of longer than 1 Sol (a Mars solar day equal 88,775 seconds) are also conspicuous, as obviously seen from the annual variation of daily averaged pressure in VL2 (Barnes 1980, 1981; Tillman 1988). It is recognized that these oscillations are probably due to the passage of baroclinic waves. Some papers (e.g. Barnes 1980, 1981; Sharman and Ryan 1980) investigated them using spectral analyses. By the analyses of Barnes (1980), there are 6–8 and around 3 Sols of dominant periodicities in surface pressure, meridional wind, zonal wind and temperature data in the autumn and spring of 1977. According to the calculation from the amplitudes of pressure, and meridional wind perturbation in the supposition of the geostrophic relation, they correspond to approximately 2 and 4 wavenumbers and $10\text{--}16\ \text{m s}^{-1}$ of the phase speeds. These wavenumbers and phase speeds calculated are generally consistent with theoretical expectations for baroclinic waves on Mars (Leovy 1969, 1979) and with the results of early Martian GCMs (Leovy and Mintz 1969; Pollack et al. 1981). In the winter of this year, soon after the “1977b” global dust storm by Tillman (1988), only the 6–8 Sols of periodicity was dominant and the phase speeds became much slower.

In studying the Martian atmospheric science, General Circulation Models (GCMs) have been used. GCMs are useful when we want to study global circulation and inspect the observational results. Comprehensive general circulation modeling of the Martian atmosphere began with Leovy and Mintz (1969), who successfully adapted the terrestrial GCM of the University of California, Los Angeles (UCLA) to Martian conditions. After this, several research groups have developed Martian GCMs and many researches have been done at NASA/Ames Research Center, French Laboratory of Dynamic Meteorology (LMD)/Atmospheric, Oceanic and Planetary Physics (AOPP) in Oxford, Geophys-

ical Fluid Dynamic Laboratory (GFDL), and so on.

The NASA/Ames Martian GCM has been used as the simulator of CO₂ polar processes and atmospheric dynamics in Mars' atmosphere (Pollack et al. 1990, 1993; Barnes et al. 1993, 1996; Haberle et al. 1993), and also used for interpretation of the Mars Pathfinder experiment (Haberle et al. 1999). LMD Martian GCM firstly simulated a full Martian year without any forcing other than insolation and topography (Hourdin et al. 1993, 1995) and was also able to reproduce the seasonal and transient pressure variations observed by the Viking Landers (Hourdin et al. 1995; Collins et al. 1996). After collaborating with AOPP, they have attempted to simulate up to the altitude of 120 km including non-LTE processes of CO₂ radiation (Forget et al. 1999; Lewis et al. 2001). GFDL Martian GCM has been used to study the role of thermal tides (Wilson and Hamilton 1996) and re-examine the Viking IRTM temperature observation (Wilson and Richardson 2000). These major Martian GCMs include surface parameters (topography, albedo and thermal inertia), CO₂ condensation-sublimation processes, CO₂ radiation codes (both thermal infrared and near-infrared absorption) and radiative effects of dust.

In this paper we present the performances of a new Martian atmospheric GCM based on a CCSR/NIES AGCM. We have introduced the important physical processes to make Martian GCM, which are CO₂ condensation-sublimation and the effect of CO₂ near-infrared absorption of solar radiation, from the GCM for terrestrial atmosphere. Moreover, we changed the surface parameters, radiative parameters of dust (refractive indices and particle size distribution) to fit the observational results of Mars. The present Martian model is described in Section 2. Numerical results and a summary are presented in Section 3 and 4, respectively.

2. Model description

The model in this study used a Center for Climate System Research (CCSR), University of Tokyo/National Institute for Environmental Studies (NIES), atmospheric general circulation model (AGCM). The basic features of CCSR/NIES AGCM are presented by Numaguti (1993), and newly implemented physical

processes are discussed by Numaguti et al. (1995). In this simulation, we changed parameters of the AGCM to adjust the environment of Mars as follows: Radius of the planet (3397 km), acceleration of gravity (3.72 m s⁻²), pure CO₂ atmospheric specific heat at constant pressure (735.9 J K⁻¹ kg⁻¹), atmospheric gas constant (188.86 J K⁻¹ kg⁻¹), solar constant (589.2 W m⁻²), component of atmosphere (CO₂: 95.3%, O₂: 0.13%), orbital elements (eccentricity: 0.0934, angle of equator inclination: 25.19°), and calendar (669 Sols a year and 88,776 seconds a day). This model does not include water processes, but includes dry convective adjustments.

2.1 Dynamics

The basic equations are three-dimensional hydrostatic primitive equations on sphere with normalized pressure (σ) coordinate. The horizontal resolution is set at about 5.6° longitude by 5.6° latitude (T21, i.e., triangular truncation with wavenumber 21). The vertical resolution is 33 layers with σ -levels, and the top of the model is about 80 km at height.

Rayleigh friction is imposed near the upper boundary. The Rayleigh friction coefficient depends on σ -level as shown in Fig. 1. The damping time near the top boundary is 0.5 day.

2.2 Surface parameters

Topography, albedo, and thermal inertia data for Mars surface are introduced. Topogra-

Fig. 1. σ -level altitude distribution of Rayleigh friction coefficient used in this model.

Fig. 2. Latitude-longitude distribution of topography used in this model [m].

Fig. 3. Latitude-longitude distribution of surface albedo data used in this model.

phy used in the model is based on the observations of the Mars Orbiter Laser Altimeter (MOLA) on the Mars Global Surveyor (MGS) (Smith et al. 1999). The data set of surface albedo and surface thermal inertia is the same as that of LMD/AOPP Martian GCM (Forget et al. 1999; thermal inertia data is fixed in Forget et al. 2001). The topography and surface albedo are averaged to the model's horizontal grid, as shown in Figs. 2 and 3. Surface thermal conductivity and heat capacity are calculated from the averaged surface thermal inertia (Fig. 4) referring to Mellon (2001) and Larsen and Dahl-Jensen (2000), and introduced in the model.

2.3 Radiation

2.3.a CO₂ gas

The treatment of absorption and emission by the strong CO₂ 15 μ m band is done by a radia-

Fig. 4. Latitude-longitude distribution of surface thermal inertia data used in this model [$\text{J m}^{-2} \text{K}^{-1} \text{s}^{-1/2}$].

tion scheme based on the k -distribution and the two-stream discrete ordinate method (Nakajima and Tanaka 1986). The scheme introduces the CO₂ radiative effects in wavenumber 550 to 770 cm^{-1} (wavelength 13.0 to 18.2 μ m) dividing 9 channels.

The effects of atmospheric heating due to absorption of solar radiation in the near-infrared band are included by calculating the heating rate by a simple analytical formula which is used in LMD/AOPP Martian GCM (Forget et al. 2003). The formula is based on the one-dimensional model calculation of Lopez-Puertas and Lopez-Valverde (1995).

In both schemes a Local Thermodynamic Equilibrium (LTE) is assumed at all altitudes as non-LTE processes are effective above the altitude of ~ 80 km (Lopez-Valverde et al. 1998), not treated in our model.

2.3.b Dust

The treatment of absorption, emission and scattering by atmospheric dust is also done by the scheme of Nakajima and Tanaka (1986). In this model, refractive indices and particle size distribution of the dust are changed to fit the up-to-date research results of Martian dust. The refractive indices data are made referring to Ockert-Bell et al. (1997) at the wavelength of less than 5 μ m, montmorillonite 219b data in Toon et al. (1977) at the wavelength of between 5 and 17 μ m, and constant values in Forget (1998) at the wavelength of more than 17 μ m. Resulting complex refractive indices are shown in Fig. 5. The particle size distribution is given

Fig. 5. Wavelength distributions of (a) Real and (b) imaginary part of refractive indices of Martian dust used in this model.

by Tomasko et al. (1999), which uses modified gamma function with an effective radius of $1.6 \mu\text{m}$ and an effective variance of 0.2, which results in the mode radius of $0.64 \mu\text{m}$, as shown in Fig. 6. These settings of the dust parameters are the same as those in the retrieval scheme used in Liu et al. (2003), which calculates dust opacity from the observational data.

2.4 Atmospheric dust distribution

We made three dust distribution scenarios which vary with time and latitude, named TES2, TES3 and VIK1 dust scenarios, to reproduce the distribution observed by MGS and Viking. TES2 dust scenario is based on the retrieved dust opacity of TES Year 2 (Year 1999) by Liu et al. (2003), and TES3 and VIK1 dust

scenarios are based on TES Year 3 (Year 2001) and Viking Year 1 (Year 1977) in the same paper, respectively. The observational results of dust opacity by Smith et al. (2001a), Smith et al. (2002) and Martin and Richardson (1993) are also taken into consideration for determining dust distributions in TES2, TES3 and VIK1 scenarios, respectively.

We used the framework of the dust scenarios of LMD/AOPP Martian GCM (Lewis et al. 2001) which calculates the optical depth of the equator, south pole and north pole by a function of aerocentric longitude L_s . When a global dust storm occurs, the increasing rate of dust opacity is rapid than the decreasing rate during decay, according to observation. To reproduce this better, we used a hyperbolic tangent transition when the dust opacity increases. The summary is as follows;

If $L_s < L_{s1}$,

$$\tau_*(L_s) = 0.05 + \frac{\tau_{\max} - 0.05}{2} \times [1 + \tanh\{(L_s - L_{s0}) * \alpha\}] \quad (1)$$

If $L_s \geq L_{s1}$,

$$\tau_*(L_s) = 0.05 + (\tau_{\max} - 0.05) \{\cos((L_s - L_{s1})/2)\}^\beta \quad (2)$$

where τ_* represents the infrared (9–10 μm) dust opacity at 7 hPa. The factors $\tau_{\max}$ is the maximum dust opacity in each regions, L_{s0} is the aerocentric longitude when dust storms occur most intensely, L_{s1} is the aerocentric longi-

Fig. 6. Particle size distribution of Martian dust used in this model.

Table 1 Values of the factors used in formulas (1)~(6) for deciding time-latitude profiles of three dust scenarios used in this paper. N, EQ and S correspond to the factors for calculating τ_N , τ_{eq} and τ_S , respectively. In VIK1 scenario two major dust storms occur, so the optical depth τ is defined by $\tau = \max[\tau_1, \tau_2]$, where τ_1 and τ_2 are the optical depths of VIK1(1) and VIK1(2), respectively, in the table.

	TES2			TES3			VIK1(1)			VIK1(2)		
	N	EQ	S	N	EQ	S	N	EQ	S	N	EQ	S
τ_{max}	0.1	0.2	0.5	0.1	0.7	1.0	0.1	0.6	0.5	0.2	1.3	1.5
L_{s0}	225°	220°	215°	210°	190°	192°	216°	206°	206°	280°	272°	272°
L_{s1}	240°	235°	235°	230°	205°	205°	230°	215°	218°	295°	283°	283°
α	12	15	8	10	18	22	12	22	22	12	18	18
β	30	30	20	20	20	15	30	10	10	30	23	23

tude when the dust opacity begins to decrease, and α and β are the factors of increasing and decreasing rates of dust opacity. Larger α represents rapider increase of dust opacity, and larger β represents faster decay. The values of these factors used in each dust scenarios are shown in Table 1. From τ_* calculated at the equator, south pole and north pole, τ_{eq} , τ_S and τ_N , the infrared optical depth at 7 hPa, τ , on any point of the planet is interpolated using a hyperbolic tangent transition in northern hemisphere and a sine transition in southern hemisphere;

$$\tau(\phi, L_s) = \tau_N + \frac{\tau_{eq} - \tau_N}{2} \times \{1 + \tanh(5(45^\circ - \phi))\} \quad (\phi > 0) \quad (3)$$

$$\tau(\phi, L_s) = \tau_S + \frac{\tau_{eq} - \tau_S}{2} \times \{1 + \sin(2(45^\circ + \phi))\} \quad (\phi < 0) \quad (4)$$

where ϕ is the latitude.

The vertical distribution of dust is calculated according to the following formula as referred in Forget et al. (1999) and Lewis et al. (1999);

$$Q = Q_0 \exp \left\{ 0.007 \left(1 - \max \left[\left(\frac{p_0}{p} \right)^{70/z_{max}}, 1 \right] \right) \right\} \quad (5)$$

$$z_{max}(\phi, L_s) = 60 + 18 \sin(L_s - 160^\circ) - 22 \sin^2 \phi \quad (6)$$

where z_{max} is the cut-off altitude of the dust [km], p is the pressure, p_0 is the standard pressure (7 hPa), and Q and Q_0 are the dust

mixing ratio at the pressure levels p and p_0 . These formulas are derived from considerations of particle sedimentation and eddy mixing by Conrath (1975), and deduced top altitude of the dust layers from the analysis by Anderson and Leovy (1978) and Jaquin et al. (1986).

The time-latitude profiles of each dust scenarios are shown in Fig. 7.

2.5 Condensation and sublimation processes of CO₂

CO₂ condensation and sublimation processes are included in this model, referring to Forget et al. (1998). The condensation temperature of CO₂ is given by the Clausius-Clapeyron relation (Hourdin et al. 1995). When the atmospheric temperature predicted from the dynamical and radiative cooling rates is below the condensation temperature, the condensation occurs and the latent heat is released to keep the atmospheric temperature to the condensation temperature. The condensed CO₂ falls to the surface immediately, then the potential energy released by ice falling and the energy used to heat the ice to the condensation temperature of lower level are taken into account in calculating the mass of condensed CO₂. The condensation on the ground is also considered as well as in the atmosphere. The loss of atmospheric mass due to condensation (or conversely the gain due to sublimation) is taken into account by modifying the surface pressure to conserve the total mass of CO₂ (caps + atmosphere). Values of surface albedo and thermal conductivity are set to 0.65 and 0.5 W m⁻¹ K⁻¹ when the surface is covered with CO₂ ice.

Fig. 7. Time-latitude profiles of the zonal-mean dust optical depth in infrared wavelength (9–10 μm) at surface in this model; (a) TES2, (b) TES3, (c) VIK1 dust scenarios.

3. Numerical results

We ran the models with the three kinds of dust scenarios described in Section 2 from northern spring equinox for three Martian years, with an initial state of isothermal conditions (240 K), a constant surface pressure of 8 hPa and no CO_2 ice. We used the third year simulation results for numerical results in the following subsection.

3.1 Distribution of temperature and wind

3.1.a Seasonal variations in the model with TES2 dust scenario

Figure 8 shows the numerical results of zonal-mean temperature in the model with TES2 dust scenario around northern autumn equinox ($L_s = 175^\circ\text{--}185^\circ$) and winter solstice ($L_s = 265^\circ\text{--}275^\circ$), respectively. Note that this figure represents the mean values of day and night time. The differences of temperature between daytime (local time ~ 1400) and nighttime (local time ~ 0200) are less than 10 K, except near the surface and above the altitude of 0.1 hPa around the point right under the sun, where the absorption of solar radiation in the CO_2 near-infrared band is the most effective. The TES retrievals correspond to a very low vertical resolution since the weighting function is very broad (Conrath et al. 2000), so the TES thermal structure may have large uncertainties, especially near the surface.

Within this context, the model results are roughly consistent with daytime observational results in TES Year 2 (Smith et al. 2003) below the altitude of 0.1 hPa, except two conscious differences. One is ~ 20 K higher temperature around the surface of the north pole at autumn equinox. This warmer autumn north pole temperature is also seen in the result of LMD/AOPP model (Forget et al. 2001). The radiometric calibration of the TES instrument may cause the errors of more than 5 K in the cold polar regions (Conrath et al. 2000). In addition to this, there may be some characteristics of the models to prevent the cooling of polar regions as the daylight time becomes shorter. The other is ~ 10 K higher temperature above southern polar regions and northern midlatitude. This seems to be because the thicker dust opacity in the southern hemisphere, and larger meridional circulation and adiabatic heating invoked by the heating by dust. Liu et al. (2003)

Fig. 8. Numerical results of time- and zonal-mean atmospheric temperature [K], (a) northern autumnal equinox ($L_s = 175^\circ\text{--}185^\circ$) and (b) northern winter solstice ($L_s = 265^\circ\text{--}275^\circ$) in the model with TES2 dust scenario. The figures are written by averaging available the same pressure-level temperature data on each grids zonally and eliminating the data at higher pressure than the zonal-averaged value. So are all the following figures which have latitude-altitude profile.

obtained significantly higher dust opacities at southern polar regions during the TES Year 2 dust storm than those of Smith et al. (2001a), who obtained dust opacities that are equal or higher in the tropics than at high latitudes. It is mentioned in Liu et al. (2003) that it is not clear whether their retrievals overestimated, those of Smith et al. (2001a) underestimated or the true opacities lie between these two estimates. Our model results may indicate the overestimate of dust opacity by Liu et al. (2003) retrievals. At the regions above 0.1 hPa, the numerical results tend to be cooler than the observations on the whole, around 20 K at 0.01 hPa except above the equator and summer hemisphere.

Figure 9 shows the numerical results of zonal wind corresponding to Fig. 8. At northern autumnal equinox the westerly wind velocity at all altitudes in the northern hemisphere is lower than the observational gradient wind shown in Smith et al. (2001a), because the latitudinal gradient of temperature shown in Fig. 8(a) is not so high as observation. At northern winter solstice the maximum easterly wind velocity reaches more than 160 m s^{-1} above southern low-latitude regions, because of higher latitudinal gradient of temperature which becomes

hotter toward south around these regions than observation. Except these points, the results are consistent with the observational gradient wind.

Figure 10 shows the mass stream function corresponding to Figs. 8 and 9. In both equinoxes the profile is near a symmetric circulation with respect to the equator, though below the altitude of 1 hPa an asymmetric pattern with respect to the equator exists, which the intensity of northern circulation is stronger than that of southern circulation. This asymmetric pattern seems to be caused dominantly by the topographic elevation difference between northern and southern hemisphere (Takahashi et al. 2003). In both solstices the characteristic meridional circulation on Mars exists, which flows across the equator from summer hemisphere to winter hemisphere. It also shows that the circulation is 2–3 times stronger in southern summer than in northern summer, which is consistent with the results of other Martian GCMs (Haberle et al. 1993; Forget et al. 1999; Richardson and Wilson 2002; Takahashi et al. 2003). The sources of this phenomenon seem to be not only the difference of the amount of insolation as mentioned in Section 1, but also the effects of Martian topography which

Fig. 9. Same as Fig. 8, except for zonal wind [m s⁻¹].

Fig. 10. Same as Fig. 8, except for mass stream function [10⁸ kg s⁻¹].

have global north-to-south elevation (Richardson and Wilson 2002).

3.1.b Interannual variability between the results of the models with TES2 and TES3 dust scenario

Figure 11 shows the daytime (local time ~ 1400) zonal-mean temperature distribution in the model with TES3 dust scenario, the difference of the daytime temperature between the

models with TES2 and TES3 dust scenarios and the difference of temperature between daytime and nighttime (local time ~ 0200) in the model with TES3 scenario, at $L_s = 205^\circ$ in the middle of planet-encircling dust storm in the models with TES3 dust scenario. Comparing them with the observational results in TES Year 3 (Smith et al. 2002), the daytime temperature below the altitude of 0.1 hPa is ~10 K larger around the equator. The difference of the

Fig. 11. Numerical results of zonal-mean thermal structure of the atmosphere at $L_s = 205^\circ$. Three days of data are averaged in each case. (a) Daytime (local time ~ 1400) atmospheric temperature [K] in the models with TES3 dust scenario. (b) The difference of daytime (local time ~ 1400) atmospheric temperatures [K] between the models with TES2 and TES3 dust scenarios, calculated by subtracting results with TES2 from results with TES3. (c) The difference of between daytime (local time ~ 1400) and nighttime (local time ~ 0200) temperatures [K] in the model with TES3 dust scenario, calculated by subtracting nighttime results from daytime results.

daytime temperature between the models with TES2 and TES3 dust scenarios is consistent above the south pole, but larger above the equator especially around the altitude of 0.1 hPa (~ 25 K). This overestimated heating seems to be due to the vertical distribution of dust, that is, these results indicate that the cut-off altitude of dust around the equator is too high in TES3 dust scenario. The global dust storm in 2001 began soon after the northern autumn equinox, earlier than the global dust storms previously observed by Mariner 9 and Viking Orbiters, when the northward meridional circulation is weaker. This may cause the lower cut-off altitude above the equator during the global dust storm in 2001, than the obser-

vational results of previous global dust storms observed by Mariner 9 and Viking Orbiters (Anderson and Leovy 1978; Jaquin et al. 1986).

The difference of temperature between daytime and nighttime in the model with TES3 dust scenario (Fig. 11(c)) is consistent with the observational result in northern hemisphere, but ~ 20 K smaller above southern midlatitude and polar region. This seems to be due to the change of dust opacity between day and night in real Martian atmosphere, which is not considered in our model.

3.2 Time evolution of CO_2 ice

Figure 12 shows the seasonal variation of surface CO_2 ice thickness in the model with

Fig. 12. Numerical results of seasonal cross-section of the zonal-mean thickness of CO_2 polar cap [kg m^{-2}] on the surface in the model with TES2 dust scenario.

TES2 dust scenario. The result of this model indicates the maximum zonal-mean ice thickness of both polar regions with $\sim 1100 \text{ kg m}^{-2}$ around spring equinoxes of both hemisphere. When we define the density of CO_2 cap 910 kg m^{-3} , the maximum zonal-mean ice thickness in this model is $\sim 1.2 \text{ m}$, consistent

with the observational results of Smith et al. (2001b).

The northern seasonal cap in this model regresses with remaining circular shape during spring, as shown in Fig. 13. But the recession speed is slower than the MOC observational results in 2000 shown by James and Cantor (2001). They indicate the averaged latitude of the edge of north seasonal cap is about 58°N at $L_s = 0^\circ$, 67°N at $L_s = 40^\circ$ and 75°N at $L_s = 70^\circ$. The TES observations in 2000 by Kieffer and Titus (2001) indicate almost the same results as the MOC observation. The latitude of the edge indicated by both the amount of seasonal cap (Fig. 13) and the ground temperature in this model is $5\text{--}10^\circ$ outside of the observational results at all these times.

During the recession of the southern seasonal cap, apparent axiasymmetrical structure of the cap is not seen in this model, even after $L_s = 240^\circ$. The averaged latitude of the edge of the south seasonal cap is about 60°S at $L_s = 219^\circ$ and 68°S at $L_s = 245^\circ$, which is about 5° outside of the observational results by Kieffer et al. (2000).

In the models with TES3 and VIK1 dust scenarios, almost the same results as in the model

Fig. 13. Numerical results of the thickness of CO_2 polar cap [kg m^{-2}] in north polar region in the model with TES2 dust scenario; (a) at $L_s = 0^\circ$ and (b) at $L_s = 40^\circ$. Solid circle in each figures indicate the latitude of 60°N .

Fig. 14. Numerical results of seasonal cross-section of the surface pressure at VG1 and VG2 in the model with VIK1 dust scenario. Because of the difference of elevation between grid points in the model and lander points (-3.43 and -2.69 km at Lander 1, -4.43 and -4.23 km at Lander 2), the values are multiplied to the raw model results by 0.93 (VG1) and 0.98 (VG2).

with TES2 dust scenario are obtained, though the thicknesses of seasonal caps in both poles are slightly (less than 50 kg m^{-2}) thicker. In these models the increase of surface temperature during southern spring and summer is suppressed by large-scale dust storms as compared with TES2 dust scenario, so the polar caps seem to grow easier at the surface.

3.3 Seasonal pressure variances

Figure 14 shows the seasonal variation of daily averaged surface pressure values at the nearest grid points of VL1 and VL2 (24.92°N , 50.62°W and 47.07°N , 225.00°W ; hereafter referred to as VG1 and VG2) in the model with VIK1 dust scenario, with modification due to the differences of elevation between grid points in the model and lander points. The pressure at VG1 varies between 6.8 and 8.7 hPa and at VG2 between 7.8 and 10.0 hPa. Comparing these results with the observational results by Viking Landers (e.g. Tillman 1988), the numerical result at VG1 is about 0.3 hPa lower at two maximum points ($L_s = 80^\circ$ and $L_s = 270^\circ$), 0.7 hPa lower at northern spring minimum point ($L_s = 350^\circ$) and consistent at northern autumn minimum point ($L_s = 170^\circ$); at VG2 consistent at two maximum points, 0.7 hPa

lower at a northern spring minimum point and 0.3 hPa higher at northern autumn minimum point. In northern spring the numerical results become relatively rather low compared to observation, probably because of the excess of northern CO_2 polar cap in this model, as mentioned in Subsection 3.2. The synoptic period oscillations attributed to baroclinic waves can be seen conspicuously between northern autumn and spring in the numerical results at both grids, with amplitudes consistent with those observed at both Landers. The synoptic period oscillations reduce significantly at both grids between $L_s = 280^\circ$ and $L_s = 310^\circ$, when the second planet-encircling dust storm is occurring, as seen in the observational results. In this period the temperature at the altitude of 0.5 hPa above VG1 and VG2 becomes very high (more than 230 K) in the model, which seems to result in high static stability of atmosphere and the reduction of baroclinic waves. Spectral analyses of these synoptic period oscillations are made in Subsection 3.5.

In the model with TES3 dust scenario the surface pressures at VG1 and VG2 become ~ 0.1 hPa lower than in the model with VIK1 dust scenario between $L_s = 190^\circ$ and $L_s = 300^\circ$, because of the increase of polar caps by the suppression of surface temperature as mentioned in Subsection 3.2. In reverse the surface pressures in the model with TES3 dust scenario become ~ 0.1 hPa higher than in the model with VIK1 dust scenario between $L_s = 320^\circ$ and $L_s = 100^\circ$, because of the suppression of surface temperature in the model with VIK1 dust scenario by the second global dust storm. In the model with TES2 dust scenario the lower surface pressures between $L_s = 190^\circ$ and $L_s = 300^\circ$ are not seen, and only the higher pressures between $L_s = 320^\circ$ and $L_s = 100^\circ$ are seen. In both the model with TES2 and TES3 dust scenarios, the reductions of oscillations between $L_s = 280^\circ$ and $L_s = 310^\circ$ are not seen.

3.4 Diurnal and semidiurnal thermal tides

Figure 15 shows the seasonal variances of the amplitudes of diurnal and semidiurnal surface pressure oscillations in the model with VIK1 dust scenario, expressed as percent of the mean pressure at VG1. This figure is made with 1-hour-mean numerical results of surface

Fig. 15. The simulated amplitudes of (a) diurnal and (b) semidiurnal surface pressure oscillations at VG1 in the model with VIK1 dust scenario, expressed as percent of the mean pressure.

pressure. The simulated amplitudes in northern spring and summer, when without global dust storms, are 0.7–1.0 percent in diurnal and 0.6–0.8 percent in semidiurnal, which are slightly smaller than observation by VL1 (e.g. Leovy and Zurek 1979). The amplitudes at VG1 become large during two global dust storms, with the maximum values of 2.4 and 2.2 percent in diurnal and 3.0 and 4.5 percent in semidiurnal. These results are 0.2–0.8 percent smaller than observation, except semidiurnal second maximum (0.5 percent larger). This shows that the global dust storms increase the amplitudes 3–6 times in the model, being consistent with observational results, though semidiurnal tide during second planet-encircling dust storm seems to be too large.

In also the model with TES2 dust scenario, which has much smaller dust opacity than VIK1 dust scenario during northern autumn and winter, the amplitudes of diurnal and semidiurnal tide increase as dust opacity increases, but the maximum amplitudes are 1.8 percent in diurnal and 1.7 percent in semidiurnal. These values are at most 2.5 times of those in northern spring and summer. In the model with TES3 dust scenario the maximum amplitudes are 2.5 percent in diurnal and 3.0 percent in semidiurnal, which are almost the same as during first dust storm in the model with VIK1 dust scenario.

3.5 Baroclinic waves

As referred in Subsection 3.3, the oscillations with the period of longer than 1 Sol in northern midlatitude winter are reproduced in our model. So we performed the spectral analyses of the numerical results of surface pressure and meridional wind at VG2 in the model with VIK1 dust scenario choosing three time periods; 1) $L_s = 210^\circ\text{--}255^\circ$ (hereafter autumn, 70 Sols), 2) $L_s = 280^\circ\text{--}313^\circ$ (hereafter winter, 54 Sols) and 3) $L_s = 325^\circ\text{--}0^\circ$ (hereafter spring, 65 Sols). The spectra are made with 1-hour-mean numerical results of surface pressure and meridional wind at the lowest layer (~ 5 m height).

In the autumn period, as shown in Fig. 16, a peak centered at the frequency (hereafter f) of 0.16 (a period of 6.4 Sols) is very evident in both variables. The second evident peak in the meridional wind spectra is seen centered at $f = 0.29$ (3.5 Sols), where a small peak is seen also in the pressure spectra. In both variables minor peaks centered at $f = 0.39$ (2.6 Sols), $f = 0.44$ (2.3 Sols) and $f = 0.50$ (2.0 Sols) are also seen. There are some additional peaks with the period of shorter than 2 Sols ($f > 0.50$) in only the meridional wind spectra.

In the winter period, as shown in Fig. 17, in the pressure spectra an evident peak is seen only centered at $f = 0.22$ (4.5 Sols). In the meridional wind spectra there are also peaks centered at $f = 0.37$ (2.7 Sols) and $f = 0.47$ (2.1 Sols), where slight peaks can be seen also in the pressure spectra, and others. The amplitudes of the peaks in both peaks become significantly smaller than in autumn and spring periods.

In the spring period, as shown in Fig. 18, peaks centered at $f = 0.18$ (5.6 Sols) and at

Fig. 16. The spectra of (a) surface pressure and (b) meridional wind at the lowest layer at VG2 in the model with VIK1 dust scenario in the autumn period ($L_s = 210^\circ\text{--}255^\circ$). The power spectral density values are given as the variance per unit angular frequency interval, where the pressure unit is hPa and the wind m s^{-1} .

$f = 0.32$ (3.1 Sols) are very evident in both variables. Other peaks seen evidently in both variables are centered at $f = 0.25$ (4.1 Sols), $f = 0.42$ (2.4 Sols), $f = 0.48$ (2.1 Sols) and $f = 0.67$ (1.5 Sols). There are also some additional peaks evident only in the meridional wind spectra with the period of shorter than 2 Sols, as well as in the autumn period. The existence of these high-frequency peaks seen only in the wind spectra would be due to the motions associated with small horizontal scales which may be nongeostrophic (Barnes 1980).

Table 2 shows planetary zonal wave numbers and phase speeds of the peaks evident

Fig. 17. Same as Fig. 16, except in the winter period ($L_s = 280^\circ\text{--}313^\circ$).

in both spectra. We calculated them with the geostrophic-balance formula (Ryan et al. 1978; Barnes 1980);

$$\lambda = 2\pi \frac{RT}{f p} \cdot \frac{p'}{v'_g} \quad (7)$$

where λ is the wavelength, T and p the mean temperature and pressure respectively, f the normal Coriolis parameter, R the gas constant, p' the amplitude of the surface pressure perturbation and v'_g the amplitude of the geostrophic meridional wind perturbation. We define v'_g using the amplitude of the meridional wind at the lowest layer v'_1 as $v'_g = 2.0v'_1$, as determined in Leovy (1979). Note that in this calculation we define the constant value of the geostrophic wind scaling factor and a turning angle of zero.

In the autumn, all the five evident waves have wavenumbers between 1.7 and 4.1, which

Fig. 18. Same as Fig. 16, except in the spring period ($L_s = 325^\circ - 0^\circ$).

is roughly consistent with the observational results of the fall period in Barnes (1980). Especially the one which is the most evident in this period (the period of 6.4 Sols) is very consistent in both wavenumber and phase speed with the 6.7-Sols-period wave of observation, which has wavenumber of 1.6 and phase speed of 14.9 m s^{-1} . Other evident waves have phase speeds between 11.3 and 24.5 m s^{-1} , the values which are consistent with or somewhat faster than observation.

In the winter, the phase speeds of all the three evident waves are between 11.0 and 13.7 m s^{-1} , which are not so slow as the observational results of winter period (5.5 and 7.9 m s^{-1}). The wavenumbers of the two minor waves are consistent with observation, though the most evident wave (the period of 4.5 Sols) has wavenumber of 2.7. In this period the mean zonal wind at the lowest layer in the model is

Table 2 Planetary zonal wavenumbers and phase speeds in the autumn ($L_s = 210^\circ - 255^\circ$), winter ($L_s = 280^\circ - 313^\circ$) and spring ($L_s = 325^\circ - 0^\circ$) periods calculated from pressure and meridional wind spectra at VG2 in the model with VIK1 dust scenario.

Time and Period [sols]	Wavenumber	Phase speed [m s^{-1}]
Autumn, 6.4	1.7	14.7
Autumn, 3.5	4.1	11.3
Autumn, 2.6	2.6	24.5
Autumn, 2.3	4.0	18.1
Autumn, 2.0	3.8	21.5
Winter, 4.5	2.7	13.7
Winter, 2.7	4.6	13.0
Winter, 2.1	7.0	11.0
Spring, 5.6	1.4	20.2
Spring, 4.1	1.5	27.5
Spring, 3.1	3.0	17.5
Spring, 2.4	3.3	20.7
Spring, 2.1	4.0	19.6
Spring, 1.5	4.2	26.0

westerly stronger than in autumn and spring, as opposed to observational results, which may result in the faster phase speeds.

In the spring, all the six evident waves have wavenumbers between 1.4 and 4.2, which are almost the same as in the autumn. The phase speeds of all the waves are more than 17.5 m s^{-1} , which are rather fast compared with the observational results. The especially evident two waves, which have the periods of 5.6 and 3.1 sols, have phase speeds which are relatively near the observation, 20.2 and 17.5 m s^{-1} , relatively.

4. Conclusions

We have constructed a new Mars climate model based on the CCSR/NIES terrestrial AGCM. To simulate the environment of Mars, we have included the realistic surface data of Mars, condensation and sublimation processes of CO_2 occurring mainly in polar regions and radiative effects of CO_2 , considering absorption of solar radiation in the near-IR bands. We also changed radiative processes of atmospheric dust with observed refractive indices and size distribution of Martian dust, and attempted to reproduce the Martian climate observed by

MGS and Viking by introducing three types of the quantitatively-consistent time- and latitude-dependent dust distribution with the retrieved observational results by Liu et al. (2003). As a result, the model can generally reproduce the distributions and seasonal variations of temperature and zonal wind observed by Mars Global Surveyor in 1999, especially at low altitude regions (below ~ 0.1 hPa). The model results of temperature tend to be lower than observation at high altitude (above ~ 0.1 hPa). We have three thinkable reasons for this. One may be that the radiative effects of water ice and vapor are not included, because in LMD/AOPP Martian GCM the temperature at high altitude is lower than observation (Forget et al. 2001). Another may be that the radiation code used in calculating the radiative effects of CO_2 strong $15\text{ }\mu\text{m}$ band is the default code of CCSR/NIES AGCM, which is made to be optimum for the troposphere of Earth, so it may have errors for the cooling rate of thin atmosphere. The third may be the lack of the effects of gravity wave drag, which would strengthen the diabatic forcing of circulation and increase the polar temperature above the altitude of 20 km in ~ 10 K (Forget et al. 1999). The numerical results of the mass stream function are consistent with the results of other Martian GCMs. The reproduction of the effects of a planet-encircling dust storm in 2001 had errors of overestimation of heating at the tropics, and underestimation of the day-night difference of temperature in southern hemisphere. To correct them, the revision of vertical distribution of dust and the consideration of the difference of day-night dust opacity may be needed.

As for the surface processes, the zonal-mean thickness of CO_2 seasonal cap in polar regions is calculated as the maximum value of ~ 1.2 m in TES2 scenario, which is consistent with the observational results by MGS-MOLA in 1999 and 2000. The edge of the northern and southern seasonal caps are outside of the observational results by MGS-MOC and MGS-TES in 1999 and 2000 during both springs. This seems to result in rather low surface pressure especially in northern spring. The shapes of regressing seasonal caps at both poles roughly remain circular until extinction, which is not consistent with the observational result at the

south pole. Further sophistication of the surface model should be expected to improve them.

The diurnal and semidiurnal tides of surface pressure become 3–6 times larger during planet-encircling dust storms than northern spring and summer at VG1 in the model with VIK1 dust scenario, which is consistent with the observational results though semidiurnal tide during the second planet-encircling dust storm is rather large. The amplitudes of the tides are 70–90% of the observation on the whole.

The synoptic period oscillations attributed to baroclinic waves can be seen conspicuously in the model between northern autumn and spring. In the model with VIK1 dust scenario the oscillations reduce significantly during the planet-encircling dust storm after the winter solstice, as seen in the observation by Viking Landers in 1977. In the spectral analyses of baroclinic waves at VG2 in the model with VIK1 dust scenario, the periods, wavenumbers and phase speeds of the waves in the autumn period are roughly consistent with the observational results by Viking Lander 2 in 1977. In the winter period during the planet-encircling dust storm, the wave with larger wavenumber than in autumn also exists, though the phase speeds of the waves are not so slow as observation. In the spring period, the phase speeds of the waves become rather fast compared with the observation.

The occurring time and intensity of dust storms during northern autumn and winter are characteristic in every Martian year, and according to them Martian climate conditions also differ in every year. Other than thermal tides and baroclinic waves observed by Viking Landers, which we inspected how the model reproduces in the model in this paper, global analyses of thermal tides and associated forced waves with the observational data by MGS have been performed (Banfield et al. 2003; Hinson et al. 2001). It will be our important subject to inspect them in our model, in order to discuss the justice of the model and improve it. We have still a lot of things to bring our Martian AGCM to be sophisticated and reliable.

Acknowledgements

The authors are grateful to Dr. Masayuki Takigawa for teaching the contents and how

to improve the CCSR/NIES AGCM, providing topographic data of Mars for our GCM and supplying much information. Also to Dr. Tatsuya Nagashima and Dr. Kengo Sudo for advise for proceeding this study, and to Dr. Naoki Sato for providing programs for spectral analyses.

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