A Simulation of Troposphere, Stratospere, and Mesosphere With an NRIJ/MA98 GCM

by

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Abstract

A preliminary result of a new atmospheric general circulation model (NRIJ/MA98) in the Meteorological Research Institute is described focusing on the seasonal and monthly mean fields. The model is based on the spectral global model, which is operationally used for weather forecasting in the Japan Meteorological Agency. The horizontal resolution is reduced to that of triangular 42 truncation (T42), but the top level is raised up to the mesopause (0.01 hPa) with an increase of layers to 45 (L45) layers. A little lower model is also made, the top of which is at 0.4 hPa just above the stratopause, with 30 layers (L30). Some physical process schemes are improved. The radiative scheme is replaced for solar and terrestrial radiation to yield sufficient accuracy in the middle atmosphere. The ground hydrological scheme is also updated. Soil layers for temperature are increased from two to three layers, similar to those for water, leading to a rigorous treatment of melting and freezing of water through the consistency between heat and water budgets.

Four- and three-year simulations are made with T42L45 and T42L30 models, respectively, under a climatological sea surface temperature, and both models reproduce reasonably well the general features of the observed atmosphere. Additional runs are performed to investigate the effects of top level position and enhanced horizontal diffusion, and their results are also described.

1. Introduction

The Earth’s climate system is composed of various subsystems such as atmosphere, ground surface, ocean and cryosphere and these subsystems are also composed of sub-systems. For example, the atmosphere includes the processes of dynamics, convection, radiation and turbulent mixing. These processes have, though there are differences in degree, strong non-linearity and affect each other through the transfer of momentum, heat and water vapor. The climate system, as a result, becomes a very much complicated system involving many non-linear feedback loops, being very difficult to investigate theoretically or with simple models. General circulation models (GCMs) contain dominant subsystems and sub-systems in themselves and thus can be a useful simulator of the climate system. To have a good GCM is, in this context, indispensable for comprehensively studying the climate system and addressing worldwide environment problems such as global warming due to anthropogenic carbon dioxide increase and ultra-violet radiation intensification due to stratospheric ozone depletion. In addition, improving GCMs leads to providing accurate medium- and long-range weather forecasts because GCMs and weather prediction models are intrinsically equivalent in this time range, though the horizontal resolution of the latter is generally much finer than that of the former.

In Meteorological Research Institute (MRI) two types of atmospheric GCMs have been used for climate study and long-range weather forecast study. One is a grided GCM, i.e., MRI GCM-7 (Tokioka et al., 1984) and GCM-II (Kita et al., 1995) and the other is a spectral GCM, i.e., MRI-GSPM (Shibata and Chiba, 1990; Chiba et al., 1995). With use of these atmospheric GCMs or ocean-atmosphere coupled
GCMs, various aspects of the atmosphere and ocean have been investigated and a number of papers have been published accordingly. For example, sea surface temperature (impact e.g. Kitooh, 1994a, 1991b; Tokioka et al., 1985), the equatorial 30-60 days oscillation (Tokioka et al., 1981), the semiannual oscillation in the southern troposphere (Yamazaki, 1980), generalized Lagrangian-mean meridional motions (Noda, 1984), penetrative cumulus convection (Dey et al., 1989), the impact of soil moisture and surface albedo changes (Yamazaki, 1989; El Nino-southern oscillation (Nagai et al., 1992), interannual and interdecadal variabilities in the Pacific (Yokimto et al., 1995), non-migrating diurnal tide (Yasui, 1989), non-parallel symmetric diurnal tide (Chiba and Siltu, 1987), and tracer transport from northern hemisphere sources (Graham and Chiba, 1990a), the seasonal and interannual variation of the stratosphere (Shibata and Chiba, 1990b), the effect of radiation scheme on the Antarctic atmosphere (Shibata and Chiba, 1990a), the solar and quasi-biennial oscillation modulation of the stratospheric circulation (Kodera et al., 1992a), the effect of the stratosphere on the troposphere (Kodera et al., 1990, 1991b) and the effect of CO2 increase (Noda and Tokioka, 1990; Tokioka et al., 1991).

In the Japan Meteorological Agency (JMA), the operational spectral global model of GSSM9811 (JMA, 1993) was upgraded in March 1996 and referred to as GSSM9903 (JMA, 1997). The highest resolution version is (triangular) truncated at wavenumber 27, corresponding to 55 km grid spacing, with 30 layers (T32L27) and is being used for medium-range forecasts of synoptic fields as well as three-day forecasts of typhoon tracks. A lower resolution version (T63L15), corresponding to L9 resolution is used for ensemble one-month forecasts. The performance of JMA models as a climate model is investigated with a reduced resolution model of T42L21 and reported by Sugi et al. (1995a, 1996b). Other features in JMA model simulations have been also investigated. For example, SST-forced variability (Sugi et al., 1997), tropical intra-seasonal oscillation (Kar et al., 1997) and Indian summer monsoon (Kar et al., 1996).

A project of making a new atmospheric GCM began in 1997 by collaborating with JMA. That is, the JMA operational global model is merged with the GCMs in MRL, retaining the highly efficient program code of the JMA model as much as possible. To be specific, on the framework of the dynamical process of the JMA model, some physical processes are replaced with those of MIPI models or improved and new ones are added. Two major changes are made for the schemes of radiation and ground hydrology. The radiative scheme is replaced with a multi-parameter random model for terrestrial radiation and with a two-stream approximation for solar radiation to maintain good accuracy not only in the troposphere, but also in the middle atmosphere. The ground hydrology scheme is improved so that the soil layers for temperature are increased from two to three layers, similar to soil water treatment, hence, the melting and freezing processes of soil water come to be rigorously treated.

The model horizontal resolution is 4, as a standard resolution, set at T112 (about 2.8 by 2.8 degrees in longitude and latitude); and the top level is raised and laid in the mesosphere. The standard model uses a 0.4 hPa top level with 30 levels. Another model for the study of the middle atmosphere, as well as the troposphere, has its top at 0.01 hPa (~80 km) around the mesosphere with 45 levels. Similar or higher top GCMs are used in various institutes/universities/centers such as the Geophysical Fluid Dynamics Laboratory (e.g. Hamilton et al., 1995), National Center for Atmospheric Research (e.g. Boville, 1995), York and Toronto Universities and Canadian Climate Center (Begley et al., 1997), Max Planck Institute (e.g. Meehl et al., 1997), United Kingdom Meteorological Office (e.g. Swinbank et al., 1996), Center National de Recherches Meteorologiques (e.g. Amodei, 1998), University of California at Los Angeles (e.g. Farrara et al., 1997) and Berlin University (e.g. Pawson et al., 1991). The intercomparison of such middle atmosphere GCMs has been made in one subproject GRIPS (GCM Reality Intercomparison Project for SPARC) under SPARC (Stratospheric Processes And their Role on Climate) project (SPARC, 1997). There are, of course, other middle atmosphere GCMs not participating in GRIPS. For example, GCMs of Kyoto University (e.g. Miyahara et al., 1993) and Center for Climate System Research of the University of Tokyo (e.g. Takahashi, 1996).

This paper reports quickly the performance of the new atmospheric GCM (324 MJTAM981-GCM), concentrating on the average fields from preliminary integrations over four years. It also contains the effects of top level position and enhanced horizontal diffusion. Detailed reports will appear in other papers based on longer time integrations, for example, the simulations for the Atmospheric Model Intercomparison Project II (AMIP II) and, coupled with an ocean GCM, the simulations for the Intergovernmental Panel for Climate Change (IPCC) report.

A brief description of the dynamical feature and physical processes of the model is given in Section 2. Model climatology for troposphere and zonal mean
structure for the entire vertical extent (1000-0.01 hPa) is described in Section 3 and stratospheric climatology is given in Section 4. The effects of changing top level height and enhanced horizontal diffusion are dealt in Sections 5 and 6, respectively and the summary appears in Section 7.

2. Model Description

The details of the dynamical features of JMA and MRI models and their physical processes are described in other papers (JMA, 1997; Kitoh et al., 1995; Chiba et al., 1996) and references therein, so that only a brief description is given here.

a. Dynamical feature

The model is a full primitive equation model and uses a spectral transform method, in which waves are trapezoidally truncated at a maximum total wavenumber 42 (T42), corresponding to the associated Gaussian grids of 128 x 64 in longitude and latitude spaced about 2.8° (~300 km). Prognostic variables are vorticity, divergence, temperature, specific humidity and surface pressure. Note that specific humidity, similar to other variables, is calculated with the spectral transform method, while it is calculated with a finite difference method in the other spectral GCM in MRI (Chiba et al., 1996). The vertical coordinate is a sigma-pressure hybrid coordinate (Simmons and Burridge, 1980), which is terrain-following in the troposphere and becomes a pressure coordinate in the middle atmosphere. The vertical level configuration is shown in Fig. 1 for 35- and 45-layer models. The 45-layer (L45) model (top 0.01 hPa) has 16 layers, which gradually thicken with height, in the troposphere (below 100 hPa) and 29 layers (210 m thickness except for the uppermost two layers) in the middle atmosphere. The half level on which the eta surface

Fig. 1. Vertical configuration of sigma-pressure hybrid coordinate for L45 (left) and L30 (right) models. Dotted and solid lines represent the left and half levels, respectively.
concludes with the pressure surface at 63.3 hPa (16th level oriented upward from the surface), i.e., $$\sigma = P$$. The 38-layer (L30) model (top 0.1 hPa) has the same layer as the 45-layer model below 10 hPa and six layers above.

Time integration is made with the semi-implicit method (Hoskins and Simmons, 1975) and uses a weak time filter (Asselin, 1972). Timestep is determined automatically in the model not to violate the Courant-Friedrich-Lewy condition and, at once, to be the greatest common divisor of a specified hour interval to synchronize the hourly time. When the timestep changes, the first time integration deviates from the normal leap-frog scheme as shown in Fig. 2 to assure it from the next timestep.

Biharmonic horizontal diffusion is used with a coefficient of $0.78 \times 10^{16}$ $m^3 s^{-1}$, which gives an e-folding time of 18 hours at a maximum total wavenumber 42. In addition, for vorticity and specific humidity, an enhanced horizontal diffusion (Simmons and Jarraud, 1983) can be switched on, which damps spectral components exceeding a critical total wavenumber dependent on the maximum wind speed for a particular model level. Through this enhancement, the advection terms for vorticity and specific humidity can be treated with the semi-implicit method, leading to a longer timestep than otherwise. As will be shown later, however, this enhancement deteriorates model climate in a certain atmospheric condition.

b. Moist convection

The Arakawa-Schubert (1974) scheme is used for deep convection with a prognostic closure similar to that of Randall and Pan (1991). The vertical profile of upward mass flux is simplified to be a linear function of height, following Mowerth and Siwee (1992). Thermodynamical properties of the upward mass flux $\sigma$ in the cloud base, which is fixed at the level near 958 hPa, are determined by the vertical mean of the grid-scale values below the cloud base. The effect of wind stress responsible for the vertical flux of the cumulus is taken into account and so is the convective downdraft, which affects the environment by decreasing the net upward mass flux and detrainsent of downdraft in the subcloud layer.

Mid-level convection, which initiates in the free atmosphere and plays an important role on the vertical mixing of heat and moisture in the stratospheres, is included in the model. Cloud forms at the level of the highest moist static energy in the vertical column and its mass flux is determined under the condition that the large scale moisture increase is spent by the convection.

c. Radiative process

Multi-parameter random model is used for terrestrial radiation as in MRI-GSM (Chiba et al., 1986; Shibata and Chiba, 1990). There are four wavelength intervals, 25-50, 50-800, 800-1250 and 1250-2200 cm$^{-1}$ and five gases can be included. H2O is treated in all the intervals with continuum absorption, while CO$_2$ (15 $\mu$m) and O$_3$ (9.6 $\mu$m) are included in the second and third wavelength interval, respectively (Shibata and Anoki, 1989). In addition to these three gases, N$_2$O (7.6 $\mu$m) and CH$_4$ (17.6 $\mu$m) are incorporated in the fourth wavelength interval. Parameters for N$_2$O and CH$_4$ are tabulated in the appendix. Full- and half-level temperatures are calculated from the prognostic lower-mean temperature profile (Shibata and Uchida, 1996) and a two-grid noise suppressing scheme is included in the integration.

Fig. 2: Schematic representation of leapfrog scheme for timestep change. If the timestep increases by 50% from $\Delta t$ to $1.5 \Delta t$ at time $P$, then the new timestep related curve is not $1.5 \Delta t$ but $1.5 \Delta t/0.5 \Delta t$ and the stability term also remains to be evaluated not at the center but at $P$. The following timesteps are evaluated by the normal leapfrog scheme.
of transmission function (Shihara, 1989). Clouds are treated as gray bodies and their optical depths are set to be proportional to the cloud water path.

Deltatwo-stream approximation is used for solar radiation. Spectral intervals, the data for k-distribution and absorption cross section and optical constants of clouds are taken from Bridgman (1962), in which taken into account are O_3 in ultraviolet and visible region (18 intervals in 0.2-8.7 µm), H_2O in near-infrared region (7-k interval in 0.3-5.0 µm), CO_2 (2.7 and 4.3 µm) and O_3 (A and B bands). Deltatwo-stream calculation for transmission and reflection is made with the discrete ordinate method (Yoshimura and Uchiyama, 1992).

To be applicable not only to the present climate simulation but also to the paleoclimate simulation, solar declination is calculated from the orbital parameters as a specified epoch, i.e., eccentricity, obliquity, longitude of perihelion, mean motion and mean anomaly, which give the sun-earth distance and the true anomaly as functions of time. In addition, the differences between the apparent and mean solar times are taken into account.

d. Ground hydrology
In the Simple Biosphere (SIB) (Sellers et al., 1986; Sato et al., 1985) model, there are three soil layers for water as shown in Table I and two soil layers for temperature, which is represented in terms of the energy storage method (Doomer, 1978). In a new version, the soil layers are intermixed from two to three layers for temperature and the diffusion equation of temperature is formulated, being the same treatment for soil water. Thus, temperature and water are defined in the same layer, so that heat and water budgets become consistent with each other. The phase transition of water, melting and freezing as a result, can be rigorously treated and this treatment possibly gives a crucial influence on the soil and near-surface conditions at high latitudes, particularly in the northern hemisphere in spring and autumn.

e. Vertical diffusion
The level 2 turbulence closure scheme by Mellor and Yamada (1974) is used with the mixing length of Blackadar (1962), in which the asymptotic mixing length is set at 300 m. Surface flux is represented with the bulk method and the drag coefficients are based on those by Lewis et al. (1982). The roughness parameters are determined from the vegetation type in the SIB scheme over the land, while it is taken from Charnock (1955) over the open sea.

f. Gravity wave drag
Orographic gravity wave drag scheme developed by Kawatsu et al. of (1989) is used, in which gravity waves are partitioned into long waves (wavelength > 100 km) and short waves (wavelength < 10 km). The long waves propagate upward and deposit momentum is the middle atmosphere, while the short waves are trapped in the troposphere and exert drag there. It is common for both waves to leave momentum fluxes are exited by the orographic exaginations of the grid scale, while the difference is the form (and hence magnitude) of the orographic variance and the way of momentum deposition.

The Rayleigh friction is also used as a two-ozone gravity wave drag. It is a hyperbolic tangent profile, similar to that in GCM in National Center for Atmospheric Research (Browning and Baumhefner, 1990).

\[ v_g(P) = \frac{\text{c} \cdot \text{c}_0}{P} \times \tanh \left( \frac{\text{c} \cdot \text{c}_0 - \text{P}_0}{\text{c} \cdot \text{c}_0} \right) \]

where \( P \) is the pressure and \( \tau_a \) (= 3 days) is the e-folding time as \( P_0 = 0.1 \text{ hPa} \) and \( \text{c} \cdot \text{c}_0 = 7 \times 1.95 \).

The vertical profile is shown in Fig. 3. This friction is applied only for the middle atmosphere, where the vertical coordinate coincides with the pressure coordinate, i.e., \( P = \text{P} \).

3. Model climatology
Four- and three-year time integrations are made without the enhanced horizontal diffusion for T21xL45 and T21xL30 models using a climatological sea surface temperature (SST). Only three major gases, H_2O, CO_2 and O_3 are taken into account in terrestrial radiation because the additional warming due to N_2O and CFC in the tropical upper troposphere (see appendix) seems not to interact with convection but to yield an undesired result, i.e., tropical tropopause rise. It implies that we convective scheme, tuned for the atmosphere containing radiative heating due to H_2O, CO_2 and O_3 alone, requires to be tuned again*. CO_2 concentration is set at 345 ppmv.

Ozone profile data is taken from Wang et al. (1995) up to 0.28 hPa and above this level the COSPAR International Reference Atmosphere (CIRA) dataset (Keating and Pitts, 1987) is smoothly merged in the MRL-GEOS model (Chiba et al., 1996). The gravity wave drag due to long waves is too included because an optimum value for the source strength is not yet determined for 1x4 and 1x30 models. Initial data are taken from a preliminary integration, which was made long enough to diminish the transient response, especially in the hydrological process, arising from the

\[ * \text{ Note added in the proof:} \]

It is found that an excess in cloud water content in upper clouds is responsible for the tropical tropopause rise.
Table 1. Layer-1, Layer-2 and Layer-3 are the top, middle and bottom soil layers, respectively. Classification and SIB biome are taken from Dormaar and Seibert (1999) except for the SIB biome of Perennial land ice.

<table>
<thead>
<tr>
<th>Classification</th>
<th>SIB biome</th>
<th>Layer-1 (cm)</th>
<th>Layer-2 (cm)</th>
<th>Layer-3 (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broadleaf-evergreen trees</td>
<td>1</td>
<td>2</td>
<td>97</td>
<td>100</td>
</tr>
<tr>
<td>Broadleaf-deciduous trees</td>
<td>2</td>
<td>2</td>
<td>97</td>
<td>100</td>
</tr>
<tr>
<td>Broadleaf and needleleaf trees</td>
<td>3</td>
<td>2</td>
<td>97</td>
<td>100</td>
</tr>
<tr>
<td>Needleleaf-evergreen trees</td>
<td>4</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Needleleaf-deciduous trees</td>
<td>5</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Broadleaf trees with groundcover</td>
<td>6</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Groundcover</td>
<td>7</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Broadleaf shrubs with groundcover</td>
<td>8</td>
<td>2</td>
<td>47</td>
<td>90</td>
</tr>
<tr>
<td>Broadleaf shrubs with bare soil</td>
<td>9</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Dwarf trees and shrubs with groundcover (tundra)</td>
<td>10</td>
<td>2</td>
<td>17</td>
<td>100</td>
</tr>
<tr>
<td>No vegetation: bare soil</td>
<td>11</td>
<td>2</td>
<td>7</td>
<td>30</td>
</tr>
<tr>
<td>Broadleaf-deciduous trees with winter wheat</td>
<td>12</td>
<td>2</td>
<td>47</td>
<td>100</td>
</tr>
<tr>
<td>Perennial land ice</td>
<td>25</td>
<td>100</td>
<td>100</td>
<td>100</td>
</tr>
</tbody>
</table>

Rayleigh Friction

![Rayleigh Friction graph](image)

Fig. 3. The vertical profile of Rayleigh friction e-folding time. The friction force is a hyperbolic tangent shape given in the text.
changes of numerical schemes mentioned in the previous section.

Observed data, for circulation, used in this study is based on the three sources, because the vertical extent and/or the accuracy of the datasets are limited within certain ranges. For the atmosphere, the CIRA dataset (Elomg et al., 1990) is used, which provides mostly and zonal mean temperature, together with zonal wind calculated from gradient wind balance. For the upper stratosphere used is the dataset (1979-96) compiled by the National Meteorological Center (NMC), which is presently known as the National Centers for Environmental Prediction (NCEP). Below 10 hPa, the National Centers for Environmental Prediction—National Center for Atmospheric Research (NCEP–NCAR) Reanalysis data (1958-1996) is used. When observation data is required over the entire vertical extent, the CIRA and NCEP datasets are so merged as to be vertically consistent. Observed data for precipitation is taken from the Global Precipitation Climatology Project (GPCP; Huffman et al., 1997) dataset but this does not cover the polar caps of 67.5°.

The total period of time integration is only four years, which is too short to evaluate interannual variations, so that only monthly mean fields simulated with T42L45 model are presented for the cardinal months, January, March, May, July, September and November, together with corresponding observed fields, unless otherwise specified. The integration period, though needless to say, for climatology evaluation should be longer than 10 years and the present four-year term is hence not a sufficient length. Nevertheless, as stated in the previous section, this report is to show a quick result and this climatology is thought not to significantly deviate from that of a long enough integration except for the winter stratosphere and mesosphere.

3.1 Sea level pressure

The simulated and observed fields of monthly mean sea level pressure are shown in Figs. 4 and 5. The model generally well reproduces the seasonal march of the geophysical distribution of sea level pressure in both hemispheres, while there are some drawbacks for specific features. The simulated Aleutian Low in November is stronger than the observed one and this trend persists during the whole period of the northern winter with diminishing in March. The simulated Icelandic Low, on the other hand, shows slightly weaker than the observed one. In summer, from May to September, both the Pacific high and Atlantic high are more strongly simulated than the observed ones.

3.2 500 hPa height

The simulated and observed fields of monthly mean 500 hPa geopotential height are shown in Figs. 6 and 7. The simulated fields are similar to the observed fields except for the northern winter, when simulations planetary waves are much stronger. The trough over the far-east Siberia is too low, being in accord with the more deeply reproduced Aleutian low in the surface. This feature that the planetary wave amplitude in the Pacific is overestimated is in sharp contrast with the result simulated with the previous JMA model of T42L21 (Sugi et al., 1991), in which the planetary wave amplitude in the Pacific is underestimated. This difference can be mostly ascribed to the precipitation increase in the subtropical western Pacific based on the convective scheme change from Kuo to Arakawa-Schubert.

3.3. 200 hPa velocity potential

Figs. 8 and 9 show the simulated and observed fields of monthly mean 200 hPa velocity potential, the negative peaks of which coincide with divergence centers in the upper troposphere. The model reproduces the overall seasonal march of the positions of negative and positive extreme values and extents of the observed velocity potential. Yet the simulated negative extreme in the west Pacific is much weaker in the northern summer. In particular, the center deviates eastward about 30° from the observed one, being coincident with the fast that the simulated precipitation in the subtropical western Pacific is much weaker in the northern summer.

3.4. 200 hPa streamfunction

Figs. 10 and 11 show the simulated and observed fields of monthly mean 200 hPa streamfunction. Since streamfunction can be used as a proxy of geopotential height in extratropics through quasi-geostrophic approximation, the simulated trough over far-east Siberia in the northern winter extending eastward indicates that it is overestimated as 300 hPa geopotential height. On the other hand, the Tibetan high, which characterizes the summer circulation in the northern hemisphere in the upper troposphere, is well reproduced as for intensity and extent during July to September, although the simulated divergent field is very different from the observed one as seen in 200 hPa velocity potential.

3.5. Precipitation

The geographical distributions of simulated and observed monthly mean precipitation are shown in Figs. 12 and 13, respectively. The model produces spatially much more concentrated rain in low latitudes.
Fig. 4. Sea level pressure (hPa) calculated from a four-year run with T42L45 model. Contour interval is 4 hPa, and the areas <1000 hPa and those >1250 hPa are shaded.
Fig. 5. Sea level pressure (hPa) calculated from observations. Contour interval is 4 hPa, and the areas <1000 hPa and those >1020 hPa are shaded.
Fig. 6. Geopotential height (m) at 500 hPa calculated from a four-year run with T42L45 model. Contour interval is 100 m.
Fig. 7. Geopotential height at 500 hPa (m) calculated from observations. Contour interval is 100 m.
Fig. 8: Velocity potential (Joules $s^{-1}$) at 200 hPa calculated from a four-year run with T42L45 model. Contour interval is $2\times10^5$ $s^{-1}$ and negative regions are shaded.
Fig. 9. Velocity potential \( \left(10^3 \text{ s}^{-1}\right) \) at 200 hPa calculated from observations. Contour interval is \( 2 \times 10^3 \text{ s}^{-1}\), and negative regions are shaded.
Fig. 15 Streamfunction ($10^5$ m$^2$ s$^{-1}$) at 200 hPa calculated from a four-year run with T42L45 model. Contour interval is $2 \times 10^5$ m$^2$ s$^{-1}$. 
Fig. 11. Streamfunction (10^5 m^2 s^-1) at 200 hPa calculated from observations. Contour interval is 2 x 10^5 m^2 s^-1.
Fig. 12. Geographical distribution of precipitation (mm/day) calculated from a four-year run with T42L45 model.
Fig. 13. Geographical distribution of precipitation (mm day$^{-1}$) calculated from observations.
than observations. In other words, the simulated rain is much larger in tropical regions such as the inter-tropical convergence zone, the southern Pacific convergence zone, middle America, Amazon and central Africa. Along with this, the zonal mean distribution of Fig. 14 (b) of the simulated precipitation in tropics is larger than the observed one every month, implicitly indicating the higher frequency of strong rain in the model. In middle latitudes the model does not well reproduce the rain along the storm track in the Pacific from July to September. In the Atlantic, on the other hand, the rain associated with the baroclinic activity is qualitatively well reproduced throughout the year. On the global average, the simulated rain is about 10% larger than the observed one.

3.6 Zonal mean zonal wind

Figs. 15 and 16 show the latitude-pressure cross section of zonal mean zonal wind for model and observations, respectively. The model well reproduces the seasonal march of zonal mean zonal wind in the troposphere and stratosphere, while the model does not in the mesosphere. This is because, though the momentum deposition of the gravity waves propagating from below plays a crucial role there, it is not included in the present model. An incorporation of non-orographic gravity waves may alleviate this error. Medvedev et al. (1997) demonstrated in the Canadian middle atmosphere model that an anisotropic non-orographic gravity wave makes the polar night jet axis inclined earthwards with height as the observed one in the southern hemisphere.

The error of simulated tropical jet is much larger in the southern hemisphere than that in the northern hemisphere, leading to more upward extension of westerly wind in early summer (November) as well as to poor separations between polar night jet and subtropical jet.

3.7. Zonal mean temperature

Figs. 17 and 18 show the latitude-pressure cross section of zonal mean temperature for model and observations, respectively. The temperature in the mesosphere, consistent with the simulated wind, is poorly reproduced in the model. Otherwise, the model well reproduces the characteristics of zonal mean temperature such as the cold (200 K) polar tropopause and sharp structure of the layer stratosphere in the summer hemisphere. However, the seasonal variation of the cold tropical tropopause area, which substantially shrinks in summer to early autumn (Fig. 18), is not well reproduced (Fig. 17).

4. Stratosphere

4.1. Northern winter

Figs. 19 and 20 show the seasonal mean (DJF) geopotential height and temperature, respectively, on 10, 30 and 100 kPa for model and observations. The temperature data on 10 kPa is taken not from the NCEP—NCAR reanalysis data, but from the NCEP operationally analyzed data, because there are small scale noisy waves in the former data. Compared with observation, the simulated polar vortex is deeper and colder for all levels. The amplitude of the trough over the east Siberia to the Pacific and the ridge over the east Pacific to the west America is too large. Namely, the Aleutian High is stronger than observations, being in accord with the simulated stronger Aleutian Low in the troposphere. This difference between model and observations can be more prominently seen in temperature field, which is more sensitive to planetary waves than geopotential field. As a result, the polar vortex deviates from the observed position, in the vicinity of the North Pole, towards the area 80°N, 50-60°E. These figures reveal the reason why the simulated zonal mean zonal wind is comparative to the observed one in high latitudes in low and middle stratosphere (Fig. 15). The simulated wind speed itself is stronger than observed ones around the polar vortex, but the off-centered polar vortex produces weaker wind speed for zonal mean operation. It should be noted again that the four-year integration is not sufficiently long to smear out the effect of sudden warmings and to evaluate model climatology, particularly in the northern hemisphere, where planetary wave activity plays a crucial role for the middle atmosphere circulation.

4.2. Southern winter

Figs. 21 and 22 show the seasonal mean (JJA) geopotential height and temperature, respectively, on 10, 30 and 100 kPa for model and observations. Since the planetary wave activity is weak in the southern hemisphere, the shape of the polar vortex is nearly concentric both in model and in observations. Yet, the simulated polar vortex is too deep, yielding stronger zonal mean zonal wind (Fig. 15) and colder zonal mean temperature (Fig. 17).

4.3. Tropical area

Fig. 23 shows the time-pressure cross section of simulated zonal mean zonal wind over the equator. Though there is no jnt of the quasi-biennial oscillation as in most GCMs, the semi-annual oscillation (SAO) is simulated in the upper stratosphere, but its westerly wind in spring and
Fig. 14. Latitudinal distributions of the zonal mean precipitation (mm day⁻¹) for model (T42L45, solid line) and for observations (GPCC, dotted line). Global mean values for model (MDL) and observations (OBS) are denoted in the upper right in each figure.
Fig. 15. Latitude-pressure cross section of the zonal mean zonal wind (m s$^{-1}$) calculated from a four-year run with T42L45 model. Contour interval is 10 m s$^{-1}$ and easterly wind regions are shaded.
Fig. 16. Latitude-pressure cross section of the zonal mean zonal wind (m s\(^{-1}\)) calculated from observations. Contour interval is 10 m s\(^{-1}\) and easterly wind regions are shaded.
Fig. 17. Latitude-pressure cross section of the zonal mean temperature (K) calculated from a four-year run with T42L45 model. Contour interval is 10 K and the regions < 200 K are shaded.
Fig. 18. Latitude-pressure cross section of the zonal mean temperature (K) calculated from observations. Contour interval is 10 K and the regions < 200 K are shaded.
Fig. 19. Seasonal (DJF) average geopotential height on 10, 30 and 100 hPa for T42L45 (left panels) and observations (right panels). Contour interval is 300 m.
Fig. 30. Seasonal (DJF) average temperature in 10, 30 and 100 hPa for T42L45 (left panels) and observations (right panels). Contour interval is 3 K and the regions <210 K are shaded.
Fig 21. Seasonal (JJA) average geopotential height on 10, 30 and 100 hPa for T42L45 (left panels) and observations (right panels). Contour interval is 300 m.
Fig. 22. Seasonal (JJA) average temperature on 10, 30 and 100 hPa for T42L45S (left panels) and observations (right panels). Contour interval is 3 K and the regions <210 K are shaded.
autumn is much smaller than observed one. This feature is similar to those in the previous GCMs in MR1 (Chiba et al., 1995; Kitoh et al., 1995), indicating the insufficient representation of eastward momentum deposition due to Kelvin waves and gravity waves responsible for the westerly phase of SAO.

5. Effect of top level height

The stratosphere (about 1 hPa level) thermally separates the stratosphere and the mesosphere, thereby being a physically meaningful boundary surface. Accordingly, the top level is sometimes set at this height in GCMs, which do not pay much attention to the middle atmosphere. Along this line, the effect of the top level height near the stratosphere is investigated. The topmost six layers of L30 model are thinned so that the top level height be at 1 hPa. This model (called L30LOW model henceforth) is also run for three years. Overall features of simulated fields are nearly the same between the two models, except for the tropical upper stratosphere, where SAO is the dominant signal. L30 model reproduces SAO (Fig. 24) as a4 model, while L30LOW model yields erroneous westerly wind persisting in a narrow region (2 to 6 hPa) with strong easterly wind above (not shown). Consistent with the wind error, temperature increase occurs in the tropical upper stratosphere through thermal wind balance, indicating that solar heating deficit due to ozone induces this coolness. The assignment of ozone concentration into the model layers of L30LOW from the observed ozone distribution yields lesser concentration in the tropical upper stratosphere, where there is a weak minimum in the observed ozone distribution. Inversely speaking, an extension of the topmost layers to the mesosphere diminishes the ozone assignment error, through which a normal thermal and hence wind structures are maintained. Fig. 24 also shows the effect of upper boundary on SAO. The easterly wind of SAO is larger by about 50% in L30 model than in a4 model, while the westerly wind occupies lesser area in L30 model than in a4 model.

6. Effect of enhanced horizontal diffusion

Using T42L45 and T42L30 models, two experiment runs are made with the enhanced horizontal diffusion, which is introduced to ensure stability for strong tropospheric jets by Simmons and Farrand (1985) and enables the semi-implicit integration of longer time step than otherwise for the advection terms of vorticity and specific humidity. Since the full analysis for T42L45 model is not yet made, only the result of the T42L30 model is stated here.

Figs. 25 and 30 show 500 hPa geopotential height and 200 hPa zonal wind averaged for December, January and February for experiment run (a), control run (b), the difference between them (c) and observations (d). It is evident that the stationary wave in the northern Pacific is stronger in the experiment run than in observations, while it is weaker in the control run. This difference pattern, which is very alike to the PNA pattern in El Niño years, appears similarly in every year, indicating a significantly confidential signal.

The Pacific jet in the upper troposphere is stronger in the experiment run than in the control run. The intensified jet area in the exit region coincides with the large difference area of Plumb (1986) wave activity flux (not shown), indicating that this PNA-like pattern gets energy through barotropic instability as demonstrated by Simmons et al. (1993a). Quantitatively, the direct effect of the enhanced horizontal diffusion is too small to induce such large impacts. For example, the critical total wavenumber is usually larger than the maximum wavenumber, 42, i.e., the enhanced horizontal diffusion not working; the critical total wavenumber barely reaches the maximum wavenumber for a condition of very strong jet such as maximum wind speed 100 m s⁻¹ and timestep 25 min. When the same enhanced horizontal diffusion is used in a horizontally reduced T42L21 GCM (Sugi, 1995a) of the previous JMA model, in which Kuo (1974) convection scheme is used, there does not appear such a stationary wave intensification in the Pacific. On the other hand, when Kuo (1974) scheme is replaced by Arakawa-Schubert (1974) scheme, the intensification of the Pacific stationary wave appears as in the present case. In the latter simulation the Pacific jet strengthens through a larger precipitation in the subtropical western Pacific, indicating that the effect of enhanced horizontal diffusion is highly dependent on the strength of the jet.

A quick look for the result of T42L45 model, however, shows that the effect of the enhanced horizontal diffusion is opposite to that for T42L30 model. That is, a negative PNA pattern appears in 500 hPa height and a weakening of subtropical jet in the Pacific. It is sufficient here to point out that the effect of the enhanced horizontal diffusion is closely connected with the amplitude of the Pacific stationary wave. A detailed analysis will be made in another paper.
Fig. 54. Time-pressure cross section of the equatorial zonal mean zonal wind (m s$^{-1}$) calculated from a three-year run with T42L30 model. Contour interval is 5 m s$^{-1}$ and westerly wind regions are shaded.
Fig. 25. Seasonal (DJF) average 500 hPa geopotential height for three-year T42L30 simulation with enhanced horizontal diffusion (g), that without enhanced horizontal diffusion (h), the difference between them (e) and observations (d). Contour interval is 100 m and shading is made for the area<5200 m for (e), (h) and (d), while contour interval is 30 m and shading is made for negative value areas for (g).
Fig. 26. Seasonal (DJF) average zonal wind on 200 hPa for (a) three-sec T42L30 simulation with enhanced horizontal diffusion, (b) that without enhanced horizontal diffusion, (c) the difference between them and (d) observations. Contour interval is 10 m s\(^{-1}\) and shading is made for the areas \(>6\) m s\(^{-1}\) for (a), (b) and (d), while contour interval is 2 m s\(^{-1}\) and shading is made for the areas \(>6\) m s\(^{-1}\) for (c).
7. Summary

A quick report of the model climatology simulated with a new atmospheric general circulation model (MRSI/MA580) in Meteorological Research Institute is provided focussing on the seasonal and monthly mean fields. The model is based on the spectral global model, which is operationally used for weather forecasting in Japan Meteorological Agency. The horizontal resolution is a triangular 42 resolution (T42) and the top level is at the mesopause (0.31 hPa) with 45 layers (L45) for the lower and 40 layers (L30) for the stratosphere. Some physical process schemes are improved. A multi-parameter random model is used for terrestrial radiation to yield sufficient accuracy in the middle atmosphere and a delta-two-stream approximation is incorporated for solar radiation. Ground hydrology scheme is so updated that temperature and water are defined at the same three soil levels, leading to a rigorous treatment of melting and freezing of water through the consistency between heat and water budgets.

Simulations of three and four years are made with the models T42L30 and T42L45, respectively, under a climatological sea surface temperature. Both models are found to reproduce reasonably well the overall features of the observed atmosphere in the troposphere and stratosphere. In the mesosphere, however, the simulated fields are poorly reproduced partly because these models do not include the momentum deposition due to gravity waves propagating from below but use the frictional form parameterization, Rayleigh friction, which cannot reverse the local wind direction. To improve the model fields in the mesosphere, an anisotropic gravity wave drag scheme is required as demonstrated by Melvedev et al. (1998).

In the northern winter, the polar vortex is simulated colder than the observed one and its position is pushed from the observed position, very near to the North Pole, towards (80°N, 60°E) by about 10 degrees due to the stronger Aleutian Highs. In zonal mean pictures, thereby, the zonal wind becomes weaker and the temperature nearly the same as the observations. In the southern winter, on the other hand, the polar vortex is also simulated colder than the observed one, but its position coincides with the observed position, the South Pole, due to the very weak planetary wave activity in the southern hemisphere. As a result, the zonal mean zonal wind is much stronger and zonal mean temperature is much colder than observations.

The effect of top level position on the stratosopause is investigated by lowering the topmost level of T42L30 model from 0.4 hPa to 1 hPa (T42L30LOW). In T42L30LOW model, the ozone concentration in the tropical upper stratosphere is underestimated through the vertical interpolation, yielding less solar heating, which in turn produces colder area there. Then erroneous wind appears and persists via thermal wind balance in the tropical upper stratosphere.

The effect of the enhanced horizontal diffusion ensuring stability for strong tropospheric jets is investigated with T42L30 and T42L45 models. It is foundAT the enhanced horizontal diffusion results in a PNA-like pattern in the northern winter if the subtropical jet is strong enough, though its direct effect is quantitatively very small.

Acknowledgments

The authors are grateful to the members of the climate research department, A. Kitoh, K. Kodera, T. Ose, H. Kodera, K. Kusunoki, K. Kuroda, A. Noda, S. Yokimoto and S. Maeda for helpful discussions and valuable comments about the model development. Computations were made with a HITAC S-880 at the Meteorological Research Institute.

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1039-1033.
Tokokki, T., A. Noda, A. Kitoh, Y. Nakayama, S. Nakagawa, T. Mito, S. Yurimoto and K. Takata,


Appendix

The parameters of multi-parameter random model (MPR) for N2O and CH4 in the wavenumber interval 1200-2200 cm⁻¹ are given here. MPR is based on the Mayr-Goody random band model for the Malkmus intensity distribution (Shibata and Ackee, 1985). The functional form of transmittance (τ) is:

\[ \ln \tau = (c_i P_0^a + c_i \exp(c_i X + c_i X^2))^{1/2} - c_i P_0 + c_i WP_0 \]

\[ c_i = c_i \exp(c_i X) \]

\[ P_0 = P_0^{W_0} X \ln(WP_0), I = \ln(\gamma T_0), \]

where \( W \) is the absorber amount, \( P \) is the pressure, \( T \) is the temperature and \( T_0 = 270 \) K. The first two terms on the right-hand side of Eq. (A1) represent the contribution of randomly distributed lines inside the band, while the last term represents the wing contribution of lines outside the band.

In calculating radiative fluxes, two types of transmission functions are required. One is a transmission function weighted by Planck function (\( B_\nu \)) and the other is the one weighted by the derivative of Planck function with respect to temperature (\( d(\nu B_\nu)/dT \)). To maintain sufficient accuracy, the parameters are determined within a certain range of pressure. For N2O and CH4, the same four pressure ranges are used as for \( H_2O, CO_2 \) and \( O_2 \). The parameters in the tables are corresponding to the following units: pressure in atmospheres, temperature in degrees Kelvin, and absorber amount in atm-cm.

The range in radiative heating due to \( N_2O \) and \( CH_4 \) for clear sky is shown in Fig. A1, in which the concentrations of both gases are 3.11 ppmv and 1.71 ppmv, respectively, and January atmospheric conditions are used. There appear two major warming (about 0.03 K/day) areas in low-latitudes and around 12 km and near surface in the troposphere, similar to past calculations (e.g. Wang et al., 1991 and Ramathan et al., 1987), while cooling (about 0.025 K/day) maximum occurs around 20 km in the summer polar stratosphere.
Fig A1. Terrestrial heating rate change due to N2O (0.311 ppmv) and CH4 (1.71 ppmv) for clear boundary conditions from the surface to 23 km. Solid and dotted lines represent heating and cooling areas, respectively. Contour interval is 0.005 K day⁻¹.

Table A1. MPR Parameters for N₂O in the Spectral Interval 1200-2200 cm⁻¹.

<table>
<thead>
<tr>
<th>c₁/c₂</th>
<th>c₁/c₃</th>
<th>c₁/c₄</th>
<th>c₄/c₅</th>
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(d/dT)Bₜ
### Table A2. MPR Parameters for N₂ in the Spectral Interval 1200-2200 cm⁻¹

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### Table A3. MPR Parameters for CH₂ in the Spectral Interval 1200-2200 cm⁻¹

\[
\left( \frac{d}{dT} \right)B_n
\]

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### Table A4. MPR Parameters for CH₂ in the Spectral Interval 1200-2200 cm⁻¹

\[
\left( \frac{d}{dT} \right)B_n
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大気ダイナミクモデル（MRI/JMA98）による対流圈、成層圏、中間圈のシミュレーション

柴田 潔央、吉村 和正、大森 淳夫、保田 敏宏、杉 正人

気象研究所の新しい共通気象モデル（MRI/JMA98）によるシミュレーションの効果的な結果を示す。月平均値に焦点を絞って述べている。モデルは気象庁の数値天気予報用の基本モデルに基づいている。水平分解能は三角形

の場合のT42に減少させたが、鉛直にはトップを中間層界（20hPa）に上げて層の数も45層（L45）に増やした。もう少し高層で、トップが成層層界の上の80hPaで30層（L30）のモデルを作った。いくつかの物理過程は

改良されている。放射過程は中層過程を中層大気でも充分に精度が高いように大気放射、地球放射スキーマと

も入れ替わられている。地球面の水蒸気過程も改良されている。中間層界の層の数を2層から3層と同じく3層

に増やし、熱とその収支を一貫させた。このため、層の混和や地表の対流の取り扱いが可能となった。

気温層の層間温度を使って、4年と3年の積分をT42,30, T42,45の2層モデルでそれぞれ行った。層モデルとも

現実大気のおのおのの特性を簡略に再現している。トップの高さの効果や強化水平伝搬の効果を調べる実験も行

い、それらの結果も述べられている。