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NIES: National Institute for Environmental Studies
CGER: Center for Environmental Research
EIC: Environmental Information Center

Foreword

The Center for Global Environmental Research (CGER), an organ of the Environment Agency's National Institute for Environmental Studies, was established in October of 1990 to contribute broadly to the scientific understanding of global changes and the elucidation of and solutions for pressing environmental problems. CGER conducts environmental researches from interdisciplinary, multi-agency, and international perspectives, provides research support facilities such as databases and a supercomputer, and offers its own data acquired through long-term environmental monitoring.

In March of 1992, CGER installed a supercomputer system (NEC SX-3, model 14) to further researches on global changes. The system is open to environmental researchers worldwide. Proposed research programs are evaluated by the Supercomputer Steering Committee which consists of leading scientists in climate modeling, atmospheric chemistry, oceanic circulation, and computer science. After project approval, authorization for system usage is provided. In 1995, two research proposals were designated as priority research and allocated larger shares of computer resources.

Volume 4 of this annual report series compiles the results of selected research activities of users of CGER's supercomputer during fiscal 1995. The papers are classified into four categories: Climate Modeling, Atmospheric and Oceanic Environment Modeling, Geophysical Fluid Dynamics, and others. The report also includes an overview of the supercomputer facilities. However this annual report does not include results from the priority researches or activities of frequent users of the Center's supercomputer which will be published in the CGER supercomputer monograph series.

We hope that this report will provide you with useful information on the global environmental researches being conducted on our supercomputer. Please do not hesitate to comment freely and directly to the Research Integration Section of CGER, so that subsequent reports will accurately reflect our research activities and effectively foster cooperation through the use of this supercomputer.

December 1996

G. Ohi

Gen Ohi
Executive Director
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Preface

The Center for Global Environmental Research (CGER) of the Environment Agency of Japan's National Institute for Environmental Studies (NIES) provides research support facilities such as a supercomputer and databases for global environmental research activities. CGER's supercomputer is open to researchers internationally for any global environmental research applications. Users need to be authorized every fiscal year. CGER is responsible for efficient allocation of supercomputer resources for each research subject, such as CPU time and memory, sufficient for the research plans recommended by the Supercomputer Steering Committee, consisting of 13 scientists (FY 1995).

NIES's Environmental Information Center (EIC) manages routine operations of the supercomputer system. This system is operated with close and cordial communications between users and the managing staff including daily consultation by the system engineers.

Progress in selected research projects is presented at an annual workshop, which also offers opportunities for communication among users and discussion of the results obtained with scientists of the global environmental area.

This report is the fourth publication of activities by the supercomputer facility, and aims to describe selected research progress from FY 1995. We hope it contributes to further progress in global change research and efforts for global environmental conservation.

December 1996

Yoshifumi Yasuoka
Director
Center for Global Environmental Research
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Contents

1. Climate Modeling

Ultra-high resolution modeling of the tropical air sea interaction: Atmospheric part

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Keywords cumulus convection, tropical atmosphere, numerical modeling

1.Backgrounds

Importance of cumulus convection in the energy and hydrological cycle of the earth's atmosphere can never be over-emphasized. But its smallnesses in temporal and spatial scales compared to the scale of the global atmosphere as a whole prevent it from explicitly calculated in current generation of GCM's. Thus every GCM includes cumulus parameterization, which have many uncertainties that arise from our poor understandings on the interaction between large scale motions and the cumulus clouds. To examine the interaction explicitly, so called 'cumulus ensemble models'(CEM) had been used since early 1980's. They are large domain (O(100km)) cloud convection models, where one imposes an 'observed' large-scale forcing and analyzes the results in a statistical manner. There remains, however, a critical defect in this approach; in the real atmosphere the interaction between cumulus clouds and large-scale motion is two-way, whereas in CEM it is forced to be one-way (large-scale motion causes cumulus convection).

Similar concern also applies to the tropical air sea interation (TASI). Usually, the TASI is thought of as a large-scale problem; typical example is El Nino and Souther Oscillation, whose spatial and temporal scales are O(10,000 km) and O(2 years), respectively. Such treatment can partially be justifiable on the ground that (1) tropical wind system that controls the air-sea exchange processes has great horizontal scale, and (2) thermal adjustment time of ocean mixed layer as a whole is long. However, atmospheric forcing to the ocean undoubtedly contains considerable smaller scale signals which is produced by cumulus convection. The significance of the small scale components to the TASI problem as a whole is still unclarified.

As documented in Nakajima(1994a), I constructed a two-dimensional cloud model that covers a domain of 16384km. Its high spatial resolution (2km) and incorporation of cloud microphysical processes allows cumulus convection to be represented explicitly. At the same time, the domain is as large as the longitudinal size of the pacific. As remarked in Nakajima(1995), if we make an ocean numerical model whose resolution and domain size is similar to the cloud model and conduct a coupled experiments, lots of informations on the TASI in the wide range of resolution will be collected. Such ocean model is still under construction, so the results can not be presented here. Instead, I present supplementaly results of the atmosphere-only experiments.

2. Combined instability of CISK and WISHE

Cumulus cloud and large scale wave disturbance can interact through wind induced modulation of surface heat and moisture flux. This mechanism is named WISHE (Wind Induced Surface Heat Exchange) by Yano and Emanuel(1991). As reported in Nakajima(1993), this mechanism produces O(10000km) scale modulation of cloud activity, and has been considered as a possible candidate of the tropical 40-60 day oscillations. Cumulus cloud and large scale wave disturbance can also interact through the direct effect of vertical air motion. This mechanism is named wave-CISK (Conditional Instability of the Second Kind). As reported in Nakajima(1994), this mechanism produces O(3000km) scale modulation of cloud activity. This may be connected to observed modulation of cloud activity such as reported by Takayabu(1993).

The situation in the real tropical atmosphere permits both WISHE and wave-CISK, so that observed cloud modulations may result from the combination of both mechanisms above. So, I examine such situations here with the following setup. Model domain size is 4096km, and the effect of the earth's rotation is omitted for simplicity. Three experiments are performed; Case W prefers WISHE but not CISK, Case C prefers CISK but not WISHE, and Case CW prefers both CISK and WISHE. The details are described below.

3. Results

In the case W, the effect of background easterly wind is introduced in the surface flux calculations. This will result in the amplification of eastward propagation of WISHE disturbance. Time evolution of the rainfall distribution in the model is shown in Fig.1. As shown there, occurrence of convective precipitation is strongly modulated in a wave number 1 propagating wave feature. This is typical behavior of WISHE disturbance in this type of cumulus resolving model.

In case C, the cooling in te model which simulates the effect of radiation is modified so as to excite wave-CISK; the cooling is enhanced in the upper half of the model troposphere; mean wind is not incorporated in the surface flux calculation, so that WISHE is excluded. Spatiotemporal pattern of the calculated rainfall is shown in Fig.2. As time goes on, the precipitation becomes more and more concentrated in a standing wave-like pattern.

In another view, one can also say that precipitation occurs as two wave disturbances, one of which propagates eastward and the other one goes west. This is typical behavior of wave-CISK disturbance in this type of cumulus resolving model.

In case CW, the wave-CISK preferable cooling is used, and WISHE preferable surface basic wind is applied, so that both mechanisms can operate. One question is whether they operate independently or some combined manner. The calculated rainfall is shown in Fig.3. It is quite evident that precipitation is concentrated in a eastward propagating patten. This behavior is very similar to that in case W.

From the result above, one might be inclined to conclude that WISHE is more 'robust' than wave-CISK. However, as described below, this is not necessarily true. Fig.4,5, and 6 show the time averaged spatial structure of the propagating wave disturbances in the cases W, C, and CW, respectively (Only figures for horizontal wind are presented. Readers are cautioned that the frame of reference for case C is chosen to move in reverse direction than the other two, so that the the sign of U is also reverse.). There, one can see that the wavey structure in the case CW is more similar to that in the case C than to that in the case W. In other words, contrally to the impression in the spatiotemporal precipitation patterns, wave-CISK is more robust than WISHE in this respect. But, it should be remarked that, although the disturbance structure is conserved, WISHE works in a very important way: it provides a strong selection rule for the direction of wave propagation.

4.Concluding remarks

Although not presented here, non wave-CISK preferable cooling was also examined. The results indicate that the cooling profile itself is important in the WISHE-CISK combined instability irrespective to whether it supports wave-CISK or not. For example, propagation speed of the large scale disturbance as well fine structure of the precipitation are sensitive to the specification of the cooling. This tendency is evident also in the large-domain experiments (not shown here).

The results of this survey implies that both of the wave-CISK and WISHE are important in understanding of the convection in the real atmosphere. It is also implied to affect to the nature of TASI, especially when one considers full range of scales. Thus, this will be one of the focuses of future coupled experiments.

Acknowledgements

Computations are done by SX-3 at CGER, NIES. The author very much thanks to the staff members of NIES for the technical supports.

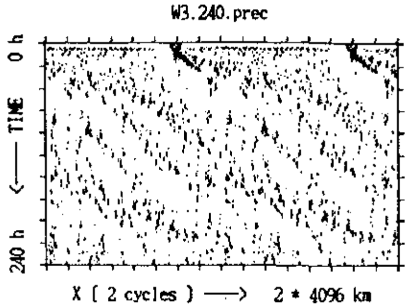

Figure 1: Time evolution of rainfall distribution in case W. Two cycles of 4,096 km domain is shown from left to right. Time goes down from 0 h to 240 h.

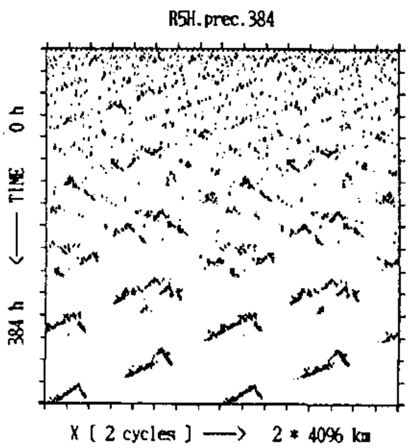

Figure 2: Time evolution of rainfall distribution in case C. Two cycles of 4,096 km domain is shown from left to right. Time goes down from 0 h to 378 h.

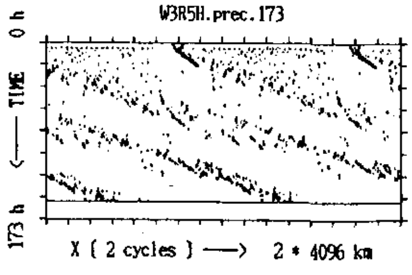

Figure 3: Time evolution of rainfall distribution in case CW. Two cycles of 4,096 km domain is shown from left to right. Time goes down from 0 h to 173 h.

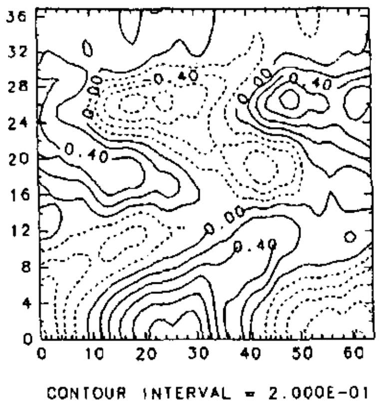

Figure 4: Time averaged disturbance wind in case W. All of the x-z domain is shown.

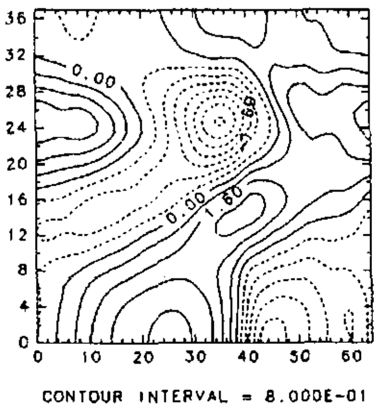

Figure 5: Time averaged disturbance wind in case C. All of the x-z domain is shown.

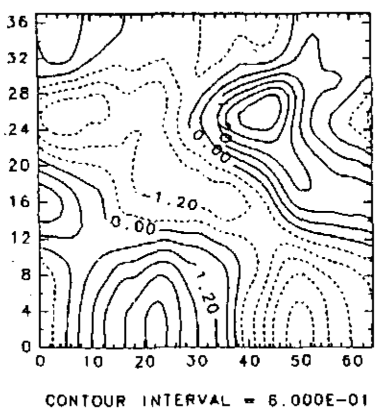

Figure 6: Time averaged disturbance wind in case CW. All of the x-z domain is shown.

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A Climate Sensitivity Study by Use of a CGCM
— On the Oceanic Residual Circulation —

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Keywords: *climate model, ocean circulation, mass circulation*

1. Background

Recent study using coupled ocean-atmosphere general circulation models (CGCM) has shown that deep ocean circulations play important roles on decadal scale climate variations and climate changes such as global warming. More recently, it has been found that the Deacon cell in the Southern Ocean cannot be regarded as a mean meridional mass circulation cell.[1-3] A similar fact has been well known in the atmospheric meridional circulation as the Ferrel cell. Thus it would be interesting to see how well the methods used for the study of atmospheric meridional circulation can be applied to the Deacon cell.

2. Objective

In a new version of the ocean part of the Meteorological Research Institute (MRI) CGCM, the formulation of the eddy diffusion tensor based on horizontal and vertical mixing has been replaced by that based on isopycnal and cross-isopycnal mixing with the Gent-McWilliams parameterization.[4] The purpose of the present study is therefore to show the sensitivity of the mean mass circulation to the formulation of the eddy diffusion tensor. However, by calculating the residual circulation[5-6], which has been so far used for the study of the atmospheric mean meridional circulation, we can see the similarity between the Deacon cell and the Ferrel cell.

3. Method

The idea underlying the residual circulation is illustrated in Fig. 1, based on the difference between the Eulerian mean and the Lagrangian mean in the presence of waves,[7-9] where (x, y, z) and (u, v, w) are the coordinates and velocity components, respectively, θ denotes a conservative quantity, and over bar denotes the Eulerian average over x. Even if fluid particles move exactly periodically (denoted by circles as projected on a y-z plane), the Eulerian average of the eddy flux $(v'\theta')$ of θ, v and w do not vanish if $\overline{\theta}$ and the amplitude of the periodic motion are inhomogeneous. The resultant wave-induced stream function becomes $\psi = -\overline{v'\theta'}/\overline{\theta}_z$, which is no more than the eddy-induced stream function obtained for the residual circulation[5] if θ is regarded as potential temperature. This stream function is irrelevant to the net mass transport.

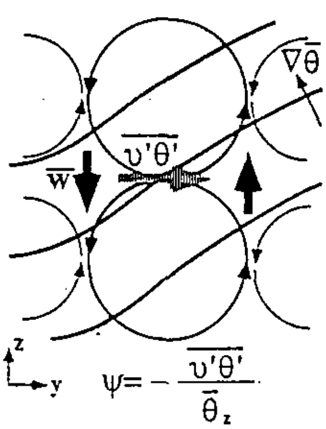

Fig. 1 An illustration of the difference of the Eulerian mean and the Lagrangian mean, and the wave-induced circulation in the presence of waves.

A figure similar to Fig.1 is found in the context of oceanic meridional circulation.[2]

4. Results

For the sake of illustration a typical atmospheric zonally averaged Eulerian mean circulation and its eddy-induced circulation in January are shown in Fig. 2. The *residual circulation* is defined as the sum of these two circulations. The data are taken from a long term integration of the MRI atmospheric GCM. The Hadley cells can be clearly seen in the latitudinal zones 40S-15S and 15S-30N, and the Ferrel cells in 65S-40S and 30N-60N. The Ferrel cells are well offset by the eddy-induced circulations in the middle latitudes where planetary wave activities are dominant.

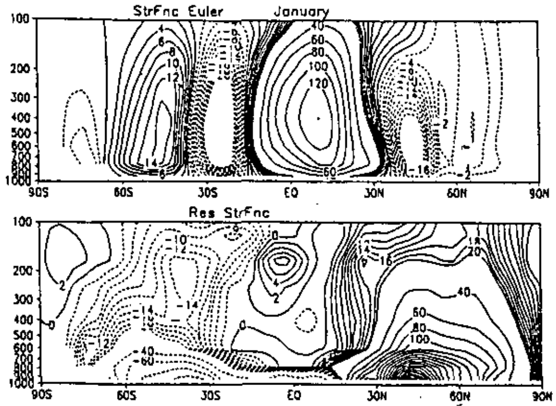

Fig. 2 A typical atmospheric zonally averaged Eulerian mean circulation (upper panel) and its eddy-induced circulation (lower panel) in January.

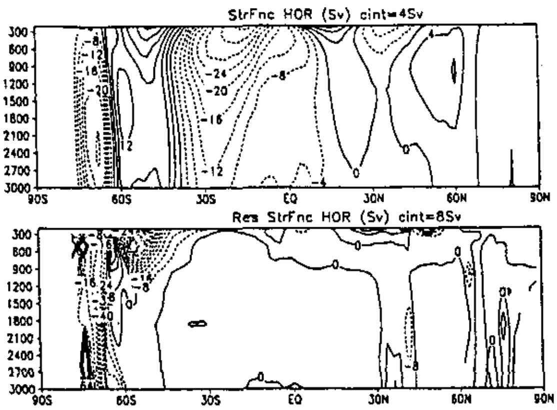

Fig. 3 An oceanic zonally averaged Eulerian mean circulation (upper panel) and its eddy-induced circulation (lower panel) obtained with an ocean model with the eddy diffusion tensor based on horizontal and vertical mixing.

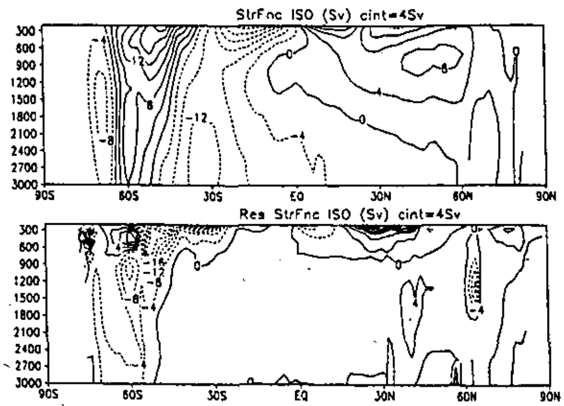

Fig. 4 As in Fig. 3 but for an ocean model with the eddy diffusion tensor based on isopycnal and cross-isopycnal mixing.

An extension to the idea to the ocean circulations can be done by taking θ as potential density. Figures 3 and 4 show the zonally averaged mass stream function and the eddy-induced stream function for an ocean model with the eddy diffusion tensor based on horizontal and vertical mixing (HOR) and for that based on isopycnal and cross-isopycnal mixing (ISO), respectively. The Deacon cell can be seen in the latitudinal zone 65S-40S in both Figs. 3 and 4. The eddy-induced circulations are dominant there. Since mixing due to the convective adjustment enhances more in HOR than in ISO, the eddy-induced circulation is not so smooth in HOR as in ISO. Thus Figs. 3 and 4 show that the Deacon cell is similar to the Ferrel cell in the atmosphere. Further implications of this similarity are under study.

Acknowledgments

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Response of the Atmospheric Angular Momentum and the Length of the Day to the Surface Temperature Increase for an Aqua Planet Model

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Keywords angular momentum, length of the day, Hadley circulation, decadal variations

1. Introduction

Various time scales exist in the variations in the Length of the Day (LOD) [1]. Among them, variations with periods shorter than interannual time scales have strong correlations with those in the atmospheric angular momentum (AAM). On the other hand, decadal variations of LOD show no apparent correlation with those of AAM derived from the recent atmospheric observed grid data. Thus, the decadal variations in LOD are thought to be caused by the fluid motions in the outer core. Nevertheless, the possibility of climatic origin of decadal LOD variations cannot be discarded easily. Lambeck and Cazenave [2] pointed out a good correlation between the decadal scale variations in the global averaged temperature and LOD (see also [3] and [4]). They found that a decrease of $\sim$ 4ms in LOD is associated with an increase of $\sim$ 0.4K in the temperature. Yoshida and Hamano [5] proposed that decadal variations of the geomagnetic field are induced by LOD changes through a topographic effect of the core-mantle boundary, and suggested that changes in LOD are caused by climate changes. They, however, did not present a mechanism how climate changes can induce decadal LOD variations.

The purpose of this study is to investigate relations between the surface temperature and AAM theoretically. Actually, it is not certain how much AAM has changed over the last several decades, since the atmospheric grid data are available only after 1970's and methods for the data archive have been continually changing [2]. Theoretical understanding is indispensable because the insufficiency of observational data prevents us to estimate from the data how much AAM should change in response to climate changes. Hence, we perform a series of GCM experiments in simplified conditions in order to estimate the response of AAM to changes in the surface temperature of the earth. We thereby obtain an estimate of the direct response of LOD to decadal changes in the surface temperature.

2. Model and Experiments

The model is CCSR/NIES AGCM [6]; the resolution is triangular truncations T42 on the sphere and 16 layers in the vertical; the Kuo parameterization scheme, simplified non-gray radiation model, and Yamada and Mellor level II diffusion model are used as physical processes. The initial state is of uniform temperature (250K) with no motion. The surface temperature is externally fixed and the surface is assumed to be everywhere wet (an aqua planet model [7]). The boundary conditions and the external forcings do not depend on time; daily and annual cycles are not included. An equilibrium state is defined as an average of 80-160 days.

We have carried out calculations under two kinds of surface temperature distribution. One represents a global temperature change. The other represents a local temperature increase. In the first group of calculations, the surface temperature is given as an annual average by

$$T_s = T_0 + (T_0 - T_1)\sin^2\varphi \qquad (1)$$

(φ is the latitude). For the standard experiment, we set $T_0 = 300$K and $T_1 = 260$K. We study dependency on the global temperature by giving a uniform deviation to T_s; $\Delta T_s = +5$K and -5K are given for the warmer and colder experiments, respectively. Figure 1 shows vertical profiles of the global mean temperature and $u\cos\varphi$, where u is the zonal wind. As the surface temperature increases, the tropopause becomes higher and the altitude of the jet (maximum of the zonal winds) also increases. The latitude of the jet is almost the same (figures are not shown). The global mean temperature profile mainly reflects that of the Hadley circulation region. The maximum of the westerlies is located at the tropopause in the polar boundary of the Hadley circulation, the extent of which is determined by the latitudinal gradient of the surface temperature rather than the absolute value of it [8].

We define U by vertical averaging of the profiles in Figure 1(b) [1]. U increases as the surface temperature increases; U = 5.9, 4.9, 4.7 m/s for $\Delta T_s = +5, 0, -5$K, respectively. This is due to the increase of the altitude of the westerly jet. This dependency is in the opposite sense to the correlation shown by [2]. In their study, since no observational winds data were available, AAM was estimated by using geostrophic winds near the surface, the winds which are balanced with the observed surface pressure distributions. The surface zonal winds of Figure 1(b) are insensitive to the surface temperature. The reason for this discrepancy to [2] may be that we have neglected the land-sea contrast and that we have relatively small surface pressure difference. If the actual land-sea contrast is taken into account, the surface pressure gradient might be enhanced. However, we have discovered the dependency of the upper level winds, which are not considered in [2]. The variation in AAM is about $\delta U = 0.1$ m/s for $\delta T_s = 0.5$K. This variation is very small: 10% of estimation of [2] and 1% of the actual LOD variation.

In the second group of calculations, we consider cases that active cumulus regions are localized as in the western pacific or in the Indian monsoon. We adopt such a surface temperature distribution that warmer deviation is localized rather than global:

$$T_s = \begin{cases} T_2 : & |\varphi - \varphi_0| < 20^\circ, |\lambda - 180^\circ| < 20^\circ, \\ T_0 + (T_0 - T_1)(\sin\varphi - \sin\varphi_0)^2 : & \text{otherwise}. \end{cases} \quad (2)$$

In these calculations, we have also studied dependency of solar seasonal inclination which is parameterized by φ_0. A localized warming T_2 is applied to the latitudes near φ_0. Figure 2 shows dependencies of U on T_2 for the various values of φ_0 (T_0 = 300K, T_1 = 260K and T_2 = 300, 305, 310K). U increases as T_2 increases in the case of $\varphi_0 < 20^\circ$, whereas U decreases as T_2 increases in the case of $\varphi_0 > 20^\circ$. In this case, since the cumulus activity is strong in the region of the maximum surface temperature, the zonal winds have a maximum (jet) near the equatorial tropopause and U is relatively large [10]. Quantitatively, the variation is $\delta U = +0.3$ m/s for $\delta T_2 = +1.0$K in the case of $\varphi_0 = 0$, and $\delta U = -0.1$ m/s for $\delta T_2 = +$ 1.0K in the case of $\varphi_0 \geq 20^\circ$. Although we have shown that U can decrease as the surface temperature increases in the case of $\varphi_0 \geq 20^\circ$, AAM variation is too small to explain the recent decrease in LOD particularly when AAM is annually averaged.

Acknowledgments

The computations were done with SX-3 at NIES. GFD-DENNOU library and GTOOL3 by Dr. Numaguti were used for drawing figures.

[1] If the sum of the angular momentum of the atmosphere and that of the solid earth is conserved, the relation between changes in the length of the day Λ ms and U m/s is given by $\delta\Lambda = 0.54\,\delta U$ [9].

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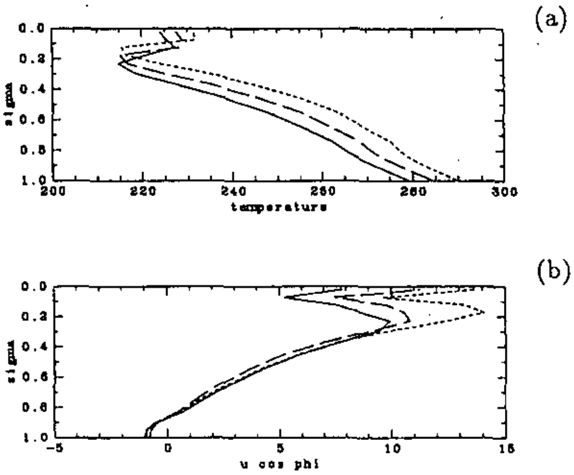

Figure 1. Vertical profiles of global mean (a) temperature and (b) $u\cos\varphi$. Solid lines: $\Delta T_s = +5$K, broken lines: $\Delta T_s = 0$K, and dotted lines: $\Delta T_s = -5$K.

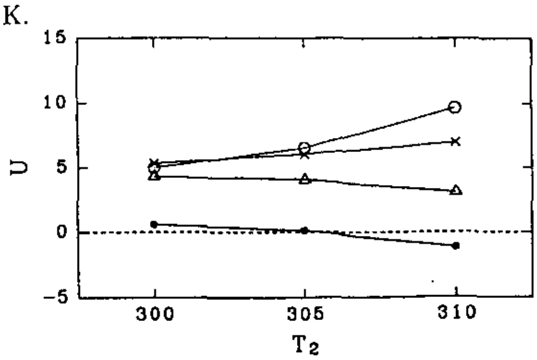

Figure 2. Dependency of U on the anomaly of tempeature T_2. Open circles: $\varphi_0 = 0^\circ$, crosses: $\varphi_0 = 10^\circ$, triangles: $\varphi_0 = 20^\circ$, closed circles: $\varphi_0 = 30^\circ$.

Mass circulation variations due to seasonal and longer term variations in the middle atmosphere circulation

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Keywords — Middle Atmosphere, Seasonal Variations, Mass Circulation, Numerical Simulation

1. Background

Air mass flows from the troposphere in low-latitudes into the middle atmosphere, where there are meridional flows from low-latitudinal region to both polar regions(Brewer, 1949; Dobson, 1956), and there occurs counter outflows from the middle atmosphere to the troposphere in mid- and high-latitudes. This indicates that the middle atmosphere plays an important role in the global circulation of trace gases such as ozone, ozone depleting chemicals, carbon dioxide. The detail of this circulation, however, is still not well understood partly because of the scantness of observations. Then model simulations are expected to be reliable tools, as a complement of observations, to investigate the global mass circulation.

2. Objective

This study is to explore the role of middle atmosphere on the global mass circulation by making and using general circulation models, which can well reproduce the characteristics of the middle atmosphere. Prior to the mass transport experiment to be made in the third year, a long term integration is made to prepare the wind dataset for off-line calculations of passive tracer transport experiment.

3. Experiment

The model used in this study is a global circulation model (GCM), which is a spectral transform model, extending from the surface to the middle mesosphere. It is a parallelogramic truncation model, *i.e.*, truncated zonally and meridionally at wavenumbers of 13 and 74, respectively, (referred to henceforth as R74), and hence has a maximum total wavenumber of 87. The horizontal resolution is about 1.4°(latitude) by 7.5°(longitude), and the vertical resolution is about 700 m with 92 layers(R74L92). The physical processes are nearly the same as in the previous GCM version (Shibata and Chiba, 1990). To avoid the noise originating from fine topography and to make stable time integration, the semi-implicit time integration is used with the reduction of the topographic resolution to be nearly equivalent with that in rhomboidal 31 truncation model. The biharmonic diffusion constant is set to be $3.25 \times 10^{14} m^4 s^{-1}$, which gives an e-folding time of about 1 day at the maximum total wavenumber 87.

4. Results

Time integration starts on June, when there is weak easterly wind in the equatorial lower stratosphere, and continues to, at present, about two and a half years. In the equatorial stratosphere there appears westerly wind around 20 hPa after 2 months or so, and becomes intensified to about 10 ms^{-1} after 6 months. Then the westerly wind gradually lowers its axis from 20 hPa to about 50 hPa during the next 9 months with slight intensification(Fig.1). Its top height, on the other hand, remains nearly constant (about 10 hPa) during this period. After the cease of the axis lowering, the top height rapidly decreases its value to about 30 hPa. After this the westerly wind shows only slight change for its extent and speed. Though the overall feature in the equatorial stratosphere is nearly the same in R13L92 and R24L92 simulations(Shibata and Chiba, 1995), the axis value of westerly wind becomes stronger with the horizontal resolution; maximum values are 15 ms^{-1}, 10 ms^{-1}, 8 ms^{-1} for R13, R24, and R74 models, respectively. Power spectral of equatorial waves, *i.e.*, Kelvin waves of wavenumber 1, 2 and Rossby-gravity waves of wavenumber 4, is slightly larger than that of R24 model(not shown), indicating that the representation of these wave may be saturated in the range between R24 and R74.

The semiannual oscillation in the equatorial upper stratosphere shows westerly phase as in the R24 model, though its vertical position of about 3 hPa is

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lower than observed one(Fig. 2). The westerly phase shows a trend of intensification with time; the value in the third year exceeds 5 ms^{-1} and 10 ms^{-1} in spring and autumn(not shown), respectively, being in contrast with the nearly constant amplitude of easterly phase. Further time integration is required to investigate this trend in detail.

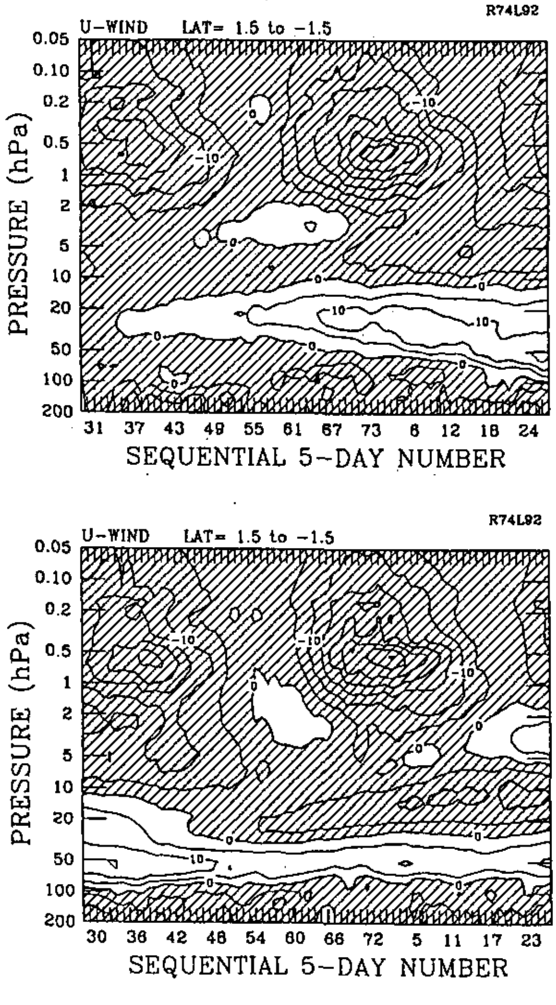

Figure.1. Time-pressure cross section of zonal mean zonal wind around the equator(1.5 °N ∼ 1.5 °S) for the first 12 months (from June to May) (upper panel) and for the next 12 months (lower panel). Contour interval is 5 ms^{-1}, and easterly wind area is shaded.

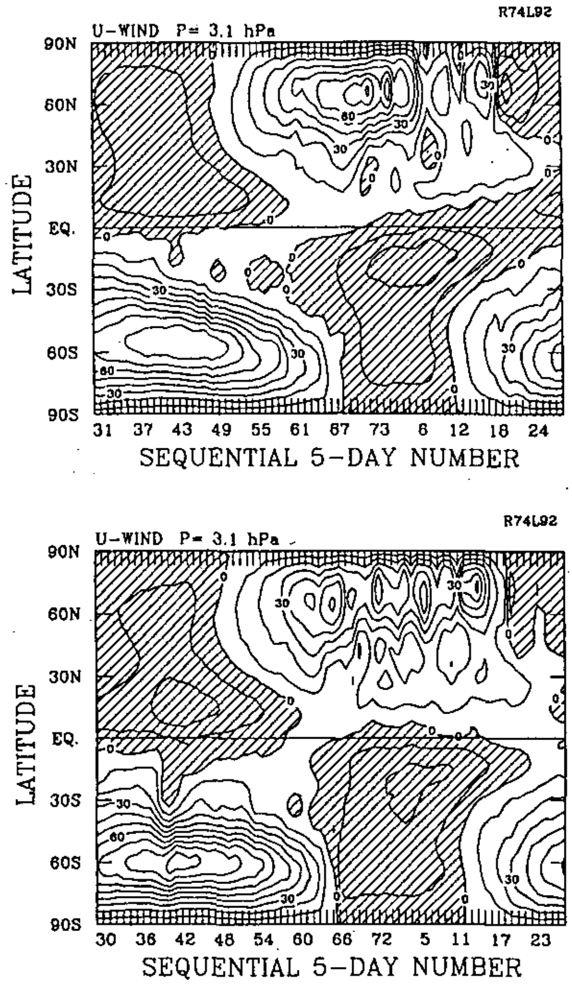

Figure.2. Time-latitude cross section of zonal mean zonal wind at 3.1 hPa level for the first 12 months (from June to May) (upper panel) and for the next 12 months (lower panel). Contour interval is 10 ms^{-1}, and easterly wind area is shaded.

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The Study of Ozone Variation with a General Circulation Model

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Key Words ozone, variation, photochemistry, general circulation model

1. Introduction

Ozone and its simulation in the climate model was discussed by H.-F. Graf and I. Yagai et al., [1995]. Current climate models incorporated various atmospheric processes that are of critical importance in the modeling of chemistry -climate interaction. Many of these seem to be handled adequately. However, some processes seem poorly handled, and require further consideration.

2. Research Objective

Vertical exchange mechanisms of ozone and precursors are very important for modeling ozone and its role in climate, and relatively poorly understood and parameterized. Stratosphere-troposphere exchange (STE) is one critical process. The exchange is controlled by fundamentally different mechanisms in tropical and extra-tropical latitudes. STE in the extra-tropics is believed to be controlled primarily through intrusions of mass in the vicinity of cyclones (tropopause folding events). These intrusions evolve to smaller scales than resolved by current global models, but the process is reasonably well represented up to the limits of their resolution, at which point the features are diffused. Current studies suggest that extratropical exchanges are modeled correctly at the qualitative, if not quantitative level.

Some of the exchange in the tropics is believed to take place through overshooting of convective turrets, or by vertical motion associated with gravity waves set up by the convection, with the balance taking place by slower, larger scale motions. The tropical exchange process is not yet well studied or understood. Some of the models which have focused on this process suggest that the model tropopause is too porous (that is the exchange takes place too easily). It is not yet clear whether this is due to inadequate vertical resolution in this region, or to errors in the formulation of convective transports. The former problem may be improved by a combination of increasing the resolution, improving the numerical formulation of those processes determining the vertical winds, or moving to a coordinate system corresponding more closely to a material surface (e.g. an isentropic coordinate). The latter problem will require a better understanding of the convective process itself.

In addition to affecting STE, it is well known that convective processes strongly influence species distribution in the troposphere itself. Recent studies suggest that both large-scale subsidence and convective downdrafts may be important in controlling tropospheric ozone.

Certain features of the stratospheric circulations are particularly critical for modeling ozone-climate interactions. Because of the temperature dependence of certain heterogeneous reactions, it is very important to get reasonable temperature distributions from GCMs. The extremely large errors-historically present in GCM simulations in winter polar lower stratosphere must be reduced. There are very important corresponding errors in winds and the mean circulation in this region. These errors are thought to exist because of insufficient wave driving by unresolved gravity waves. Current gravity wave drag formulations adopted in the MRI GCM includes only the sources from zero phase speed gravity wave excited by flow over orography, neglecting those waves with a broader spectrum of phase speeds being excited by convection, shear, and frontal propagation. Including these effects may have a very substantial and beneficial impact on polar night temperatures. Because the role of gravity waves in the atmosphere is so poorly understood there is a need for detailed modeling as well as data studies in order to improve our understanding of these processes.

3. Results

The 12-layer general circulation model of the Meteorological Research Institute (MRI GCM) is used in this study which has seven layers in the stratosphere with the top of 1 hPa. The current model roughly simulate the seasonal variation of zonal mean ozone mixing ratio (Yagai, [1995]). However, the model also has problems mentioned above and the simulated results below are preliminary results.. Figure 1 shows polar stereographic projections of the simulated monthly mean total ozone for January. and the correspond observed values are seen in Figure 2. The largest value is seen over the Sea of Okhotsk and the second largest value over Canada, which corresponds to the observations. The minimum value is seen over Green·land and the horizontal wavenumber 1 dominates at high latitudes in the Northern Hemisphere. The same quantities as in Figures 1 and 2 but for month of March, are shown in Figures 3 and 4. The large values are seen above the eastern part of Siberia and Canada in the simulation and the observation, which indicates the dominance of horizontal wavenumber 2 at high latitudes. As a result the model successfully simulates the horizontal distribution of total ozone in quality during winter to spring when the dynamical processes are dominate to determine distribution of ozone. To obtain the quantitative agreement with the observation in all seasons is the next step in this study.

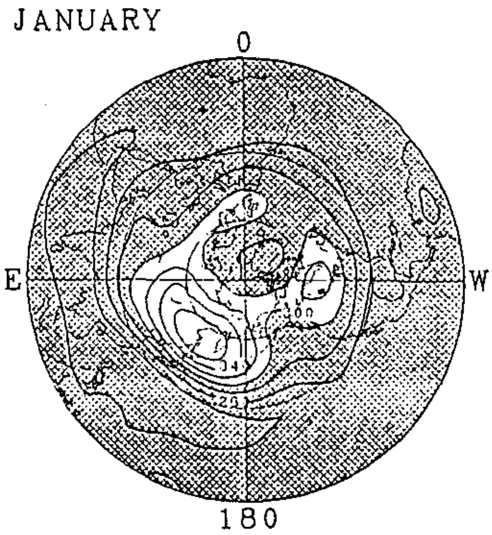

Figure 1: Horizontal map of total ozone in January from the result by the MRI GCM. The contou interval is 20 Dobson units and the values less than 300 are shaded.

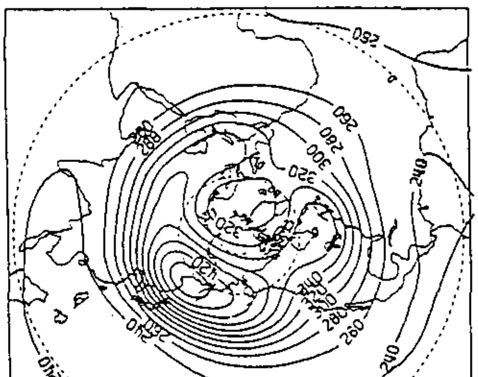

Figure 2: The 8-year average (1979-1986) total ozone distribution for January (1979-1986) [Nagatani et al., 1988] from the Nimbus 7 Solar Backscatter Ultra Violet (SBUV) instrument.

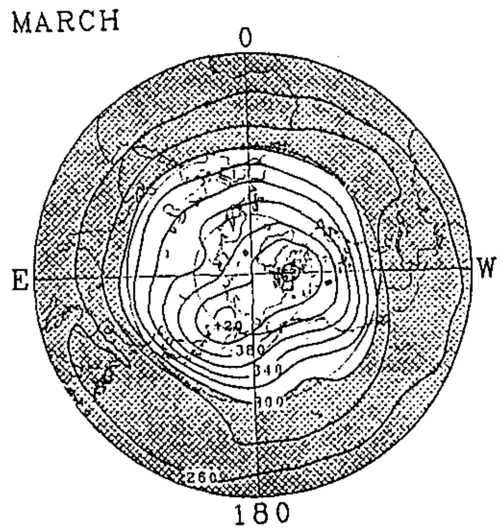

Figure 3: The same as in Figure 1 but for the month of March.

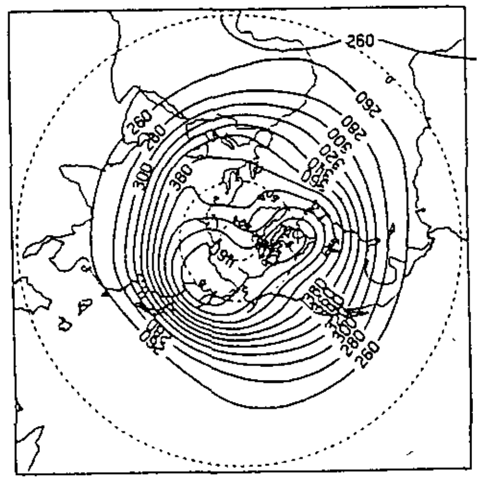

Figure 4: The same as in Figure 2 but for the month of March.

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