CGER’S SUPERCOMPUTER

ACTIVITY REPORT

Vol.12 – 2003

Center for Global Environmental Research

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Foreword

The Center for Global Environmental Research (CGER) of the National Institute for Environmental Studies was established in October 1990. The main objectives of CGER are to contribute broadly to the scientific understanding of global change and to the elucidation of and solutions for our pressing environmental problems. CGER conducts environmental research from interdisciplinary, multi-agency, and international perspectives, provides research support facilities such as a supercomputer and databases, and offers to the public its own data from long-term monitoring of the global environment.

Since March 1992, CGER installed a supercomputer system (NEC SX-3, Model 14, in 1992-1997; NEC Model SX-4/32, in 1997-2002), and we are operating NEC Model SX-6 at its full capability due to the increase of CPU utilization, and we are making efforts to operate it more effectively.

Proposed research programs are evaluated by the Supercomputer Steering Committee consisting of leading Japanese scientists who have concerns with global environment issues, such as climate modeling, atmospheric chemistry and oceanic circulation, or computer science. After project approval, authorization for system usage is provided. In fiscal year 2003 (April 2003 to March 2004), 21 research programs were approved to use the supercomputer system.

The Supercomputer Activity Report Vol.12 compiles the research results in fiscal year 2003. The research papers in this report do not necessarily show the final results of the research programs, which are to be published in the form of "full papers" upon completion of each research program. This report consists of research papers classified into four categories —Climate Modeling, Atmospheric and Oceanic Environment Modeling, Geophysical Fluid Dynamics, and Other Researches—together with the Overview of the Supercomputer Systems.

We hope this report provides useful information on the global environmental research. In order to promote the exchange of ideas and opinions amongst the scientific fraternity about utilizing this supercomputer, the Research Integration Section of CGER would greatly appreciate any comments or suggestions on this publication.

March 2005

Shuzo Nishioka
Executive Director
Center for Global Environmental Research
National Institute for Environmental Studies
Preface

The Center for Global Environmental Research (CGER) provides a supercomputer system to stimulate the global environmental research activities. It is widely open to researchers for modeling and requires application of a user. CGER grants access permission to the system with a priority rank without charge as long as the research objective is to contribute to the global environment research. The system is supported by the grant from the government, and the accountability to the people is to open research results obtained by using the system. This activity report is the minimum obligation of publishing scientific papers to contribute to the global environemnt research.

Owing to the efforts of Environmental Information Center of NIES, the supercomputer system is working without serious problems.

We hope this publication contributes further to the progress in global environmental modeling and to the global environmental conservation.

March 2005

Gen Inoue
Director
Center for Global Environmental Research
National Institute for Environmental Studies
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1. Climate Modeling
Year-to-Year Variation in Total Ozone Minimum over the Subtropical Western Pacific Region

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Abstract
The year-to-year ozone variation over the subtropical western Pacific region is studied, using the Earth Probe Total Ozone Mapping Spectrometer (EP_TOMS) ozone data from August 1996 to July 2002. Regression analyses show that dynamical signals, such as the quasi-biennial oscillation (QBO), play an important role in determining total ozone variation. The CCSR/NIES nudging chemical transport model (CTM) is used to simulate the year-to-year ozone variation and explain the mechanism for producing ozone lows in a 3-D distribution of ozone. The year-to-year ozone variation, especially the winter ozone low, is well simulated by the model excluding heterogeneous reaction processes between 45°S and 45°N latitude. Results of the model calculation including heterogeneous reactions show that the heterogeneous reaction effect is small and the year-to-year ozone variation is mainly controlled by dynamical transport processes.

Keywords: Year-to-year ozone variation, Nudging CTM, Subtropical western Pacific, QBO

1. Introduction

Kawahira et al. (2002) found that in December 2001 EP_TOMS observed total ozone less than 225 dobson unit (DU) around (20°N, 130°E) over the subtropical western Pacific. They also showed that total ozone of less than 200 DU was observed by ground based instruments at Hong Kong (22.22°N, 114°19′E), China, and Naha (26.12°N, 127°40′E), Japan, in the subtropical western Pacific for a few days in December. The lowest value observed was less than 190 DU. Such a local low ozone value stimulated us to study the mechanisms determining the ozone variation over the subtropical western Pacific. EP_TOMS observations show that the total column ozone was unusually low in the winters of 1996/97, 1998/99, and 2001/02 (see Figure 1). Thus it is useful for studying the year-to-year ozone variation in this region in order to address the following questions. Is it possible to explain such low ozone levels in the western Pacific by just the QBO and ENSO effects? Is there any possibility that heterogeneous reactions can cause such low values of ozone? Few studies on the local ozone minimum over the western subtropical region have been carried out in order to explain which process (dynamical process, or chemical process, or both) are really responsible for creating such an ozone low. Han and Yamazaki (2002) analyzed the vertical distribution of ozone in the subtropical Pacific region using the HALOE/UARS ozone data. They found that the negative deviation of total ozone from the zonal mean value was large at the pressure levels of 10-15 hPa and 40-60 hPa for the 10-year climate mean between 1991 and 2000. They suggested that the negative deviation at the lower altitude was due to the advection of low-ozone air from the tropics and that, at the higher altitude, it was due to the advection of low-ozone air from the Aleutian high.
In this paper, possible factors affecting the year-to-year ozone variation are analyzed using a regression method, and the variation is simulated using the CCSR/NIES nudging CTM. By comparing the modeling results with the observations, our objective is to investigate whether the ozone low in a 3-D distribution of ozone can be caused by dynamical processes and to discuss the possible contribution of a heterogeneous chemical process.

2. Regression Fit Analysis of EP_TOMS Data

2.1 EP_TOMS Data and Analysis Method

The total ozone data used here are from EP_TOMS for August 1996 to July 2002, and from Nimbus-7 TOMS data for 1979 to 1992. EP_TOMS provides the global ozone distribution with a resolution of 1° latitude by 1.25° longitude.

The regression methods used in this study are similar to those of Ziemke et al. (1997), but slightly different in neglecting the trend term, due to short analysis period. The regression equation has the following form:

$$TO_3(t) = \alpha + \beta \text{Solar}(t) + \gamma \text{QBO}(t-\lambda) + \delta \text{ENSO}(t) + \epsilon \text{MSU}(t) + R(t),$$

where \(\alpha, \beta, \gamma, \delta\) and \(\epsilon\) are time-dependent regression coefficients given by a constant plus 12-month, 6-month, and 4-month cosine and sine harmonic series. The MSU temperature is the microwave sounding unit channel 4 brightness temperature. Channel 4 measures the thermal radiation emitted from an atmospheric layer centered between approximately 50-150 hPa (13-22 km), with a maximum response near 90 hPa (17 km). In the equation, \(\alpha\) is the seasonal variability, and \(\beta \text{Solar}(t)\), \(\gamma \text{QBO}(t-\lambda)\), \(\delta \text{ENSO}(t)\), and \(\epsilon \text{MSU}(t)\) represent the solar, QBO, El Niño/Southern Oscillation (ENSO), and MSU temperature-associated regression fits, respectively. \(R(t)\) is the residual time series, representing the regression error. A trend term is not included in the study because it is implicitly nonlinear, causing a lack of reliability with the trend fit (Randel and Cobb, 1994). In fact, our results show that there is almost no significant trend over the subtropical region, which is consistent with the assessment of the World Meteorological Organization (WMO) (1999). The solar signal is from the 10.7 cm solar flux series, and the QBO signal represents the quasi-biennial oscillation index from Singapore (1°N, 104°E) 30-hPa zonal winds. The time series of the QBO signal are selected as in Randel et al. (1995), adjusting the phase lag to maximize the cross correlation with the QBO signal in equatorial TOMS ozone, and then using the same lag value for other latitudes. The phase lag value here is 5 months. The ENSO signal is obtained from the difference between Tahiti and Darwin-normalized sea level pressure. Prior to analysis, the time-averaged values are removed from the solar, QBO, and ENSO time series. The MSU temperature time series are deseasonalized and detrended from their monthly mean data. All the time series used here are monthly mean data, covering the period from August, 1996 to July, 2002. An area-weighted (cos[latitude]) mean is applied to the region of 15-25°N, 120-150°E, which is selected to represent the subtropical western Pacific region. Note that the solar term was neglected for the EP_TOMS data analysis between 1996 and 2002, because the inclusion of this term did not make any improvement in the result, in this case of short analysis period.
2.2 QBO, ENSO, and MSU Temperature Signals

After removing the seasonal variability from the total ozone variation, there is an ozone anomaly, which should include the solar, QBO, ENSO signals, and MSU temperature variations. The solar and ENSO-associated regression fits indicate small amplitudes, ranging from 2% to 3% in total ozone, while the QBO and MSU temperature-associated regression fits indicate contributions range from 5% to 7% to the total ozone.

![Ozone anomaly and its regression terms over the SSWP, from EP_TOMS data](image)

Figure 1  Time series of EP_TOMS deseasonalized ozone anomaly from August 1996 to July 2002 over 15°-25°N, 120°-150°E region. Total ozone anomaly, QBO anomaly, ENSO anomaly, and MSU temperature anomaly are indicated by the black, green, blue, and red lines, respectively. The unit for the vertical axis is Dobson unit (DU).

Figure 1 shows the deseasonalized ozone anomalies calculated by the regression fit. The QBO-associated regression fit (the green line) explains most of the ozone anomaly during the period except in the autumn of 2000 and the winter of 2000/2001, when the MSU temperature (the red line) makes a major contribution to the ozone anomaly. In particular, the low ozone anomalies in the winters of 1996/97, 1998/99, and 2001/02 were closely related to the QBO since the associated regression fit indicates that the anomalies are caused by the phase transitions of the QBO zonal wind during these periods. From 1996 to 2002, the equatorial zonal winds of QBO experienced three phase variations, and the transitions from the easterly wind to the westerly wind occurred around 30 hPa in the winters of 1996/97, 1998/99, and 2001/02 (Figure 2). Therefore, in the winters listed above, the vertical motion associated with the QBO is in the eastward transition phase (from easterly to westerly) above 30 hPa, which causes a sinking motion along the equator and an upwelling in the subtropics according to the near-geostrophic thermal wind balance (Andrews et al., 1987). In fact, this upwelling is clearly seen from the geopotential height anomaly of the European Centre for Medium-Range Weather Forecast (ECMWF) data (Figure 3). In this figure, the positive height anomalies prevailed at almost any altitude above 100 hPa in the winters of 1996/97, 1998/99 and
2001/02, representing an upwelling during these periods. The upward motion brings the lower ozone concentrations of lower altitudes to the higher altitudes below the ozone number density maximum height (about 20 hPa). It produces negative total ozone anomalies in the winters of 1996/97, 1998/99, and 2001/02. The correlation between QBO and ozone that we found here is consistent with previous studies (Zerefos et al., 1992; Trepte and Hitchman, 1992; Yang and Tung, 1995; Cordero, 2002).

Figure 2. Time series of zonal wind at 2.7°N and 106°E from ECMWF (the upper panel) and area-weighted mean total ozone over 15°-25°N and 120°-150°E region (the lower panel). The contour interval for the upper panel is 10 m s⁻¹, and the dotted lines represent easterly wind. The solid line and the dashed line in the lower panel are the variations from EP_TOMS and the nudging CTM with ECMWF data, respectively. The vertical red-dotted lines indicate the three low-ozone winters.
Figure 3  The time-pressure section of geopotential height anomalies over 15°-25°N, 120°-150°E region. The data are from ECMWF, and the dotted lines represent negative values. The contour interval is 2 m. The area-weighted mean values in the region are shown.

3. CTM Study

3.1 CCSR/NIES Nudging Chemical Transport Model

A nudging chemical transport model (CTM) was developed at NIES using the CCSR/NIES AGCM. The chemical scheme and chemistry-radiation coupling scheme of a 1-D coupled chemistry-radiation model developed by Akiyoshi (2000) are incorporated into the GCM. The present version includes BrOx chemistry and heterogeneous reactions on polar stratospheric clouds (PSCs) of STS/NAT/ICE as well as the Ox, HOx, NOx, hydrocarbons, and ClOx gas phase chemical reactions for the stratosphere (Akiyoshi et al., 2002). In this study, however, heterogeneous reaction processes are excluded between 45°S and 45°N to study the gas phase chemistry and transport effects on ozone variation. A horizontal resolution of T21 (longitude×latitude=5.6°×5.6°) is employed and the vertical domain extends from the surface to about 70 km with 30 layers. The zonal wind, the meridional wind, and the temperature of the model are nudged toward the ECMWF data. The original ECMWF data available to us have a resolution of 2.5°×2.5° which were then interpolated to the T21 model. A relaxation timescale of 1 day is used for the nudging procedure. Above 10 hPa, where no ECMWF data exist, the monthly temperature data and zonal wind data of the Committee on Space Research (COSPAR) International Reference Atmosphere-86 (CIRA-86) are used.
3.2 Comparison between Observation and Nudging CTM

3.2.1 Horizontal Distribution

The left panels of Figure 4 shows the global total ozone distribution calculated from the model and the TOMS ozone measurements in the winter of 1996/97. An ozone low is obvious over the subtropical western Pacific, centered at 20°N and 146°E, with values less than 230 DU. In the model, the ozone low is well simulated, with almost the same position and ozone value as the observations. The ozone high in the middle and high latitudes related to the Aleutian high is also well simulated, with the same maximum value as the measurements. The position of the Aleutian high of the model was slightly south of that of the EP_TOMS.

![Figure 4](include:image1.png)

Figure 4 Horizontal distribution of total ozone from TOMS in the winters of 1996/1997 (a), that from the nudging CTM (b), that from TOMS in the winters of 1997/1998 (c), and that of the nudging CTM (d). The contour interval is 10 DU.

In the 1997/98 winter, the ozone low disappeared over the subtropical western Pacific and moved equator-ward and eastward to the tropical central Pacific (see the right panels of Figure 4). The equator-ward displacement of the ozone low is closely related to the QBO (see discussion in section 6.2), and the eastward displacement is closely related to the extension of the sea surface temperature to the eastern Pacific during the 1997-1998 ENSO event (e.g. Chandra et al., 1998; Thompson and Hudson, 1999). The difference between the winters of 1996/97 and 1997/98 is also well simulated in our model, together with the ozone high over
the Aleutian region. Compared with the previous year, the total ozone in this winter was higher over the subtropical western Pacific.

In the other winters (not shown), the position of the ozone lows are also simulated well using the nudging model excluding the heterogeneous reaction processes between 45°S and 45°N.

3.2.2 Year-to-Year Variation in the Subtropical Western Pacific Region

The lower panel of Figure 2 shows year-to-year ozone variation for EP_TO MS and nudging CTM over the subtropical western Pacific. An area-weighted mean has been applied. A similar year-to-year variation was produced by the model except for the winter of 1999/2000, showing that the ozone gas phase chemistry and the dynamical processes played an important role in determining total ozone in this region. The model also succeeded in simulating the unusually low ozone in the winters of 1996/97, 1998/99, and 2001/02. In fact, an ozone minimum, lower than 230 DU over the Pacific region, existed for all winters except for 1999/2000. However, the position of the ozone minimum was different in different winters. The position variation along the meridian is consistent with the meridional circulation induced by QBO. For example, in the unusually low ozone winters of 1996/97, 1998/99, and 2001/02, the ozone minimum was over the western subtropical Pacific, corresponding to the western phase of the QBO signal above 30 hPa. During this period, the QBO induced an upward motion anomaly in the subtropics (Trepte and Hitchman, 1992), weakened the subsidence branch of the Hadley circulation and caused the total ozone minimum. In the other years, the ozone minimum was over the equator, corresponding to the easterly phase of the QBO, which caused a downward motion anomaly and positive ozone anomaly over the subtropics.

3.3 Possibility of Ozone Loss Due to Heterogeneous Reaction Processes

The effect of the heterogeneous reaction processes on the ozone variation could not be excluded because chlorine loading in the stratosphere reached a maximum during 1996-2002 (WMO, 1999). In the simulation of the ozone low during 1996-2002 with the CTM, in which the heterogeneous reaction processes are excluded between 45°S and 45°N, some differences were found between the model results and the observations. For example, the EP_TO MS ozone minimum value in the winter of 1998/99 was lower than that in the winter of 1999/2000, but the reverse relation occurred in the nudging CTM results (the lower panel of Fig. 2). The discrepancies might be due to the heterogeneous reaction processes. For further discussion, results from another model run are provided here to check the importance of heterogeneous reaction effects. The formation of nitric acid trihydrate (NAT) and ice were allowed over the tropics and subtropics in the model run. The condensed volume was estimated by the difference between the partial pressure of H₂O, HNO₃ and their saturation vapor pressures. Hence, the saturation ratio for condensation was set to be unity. The volume mixing ratios of H₂O and HNO₃ in the lower stratosphere in the model are around 2 ppmv and 2-3 ppbv, respectively, over the subtropical western Pacific in winter. These are consistent with observations (e.g. Kumer et al., 1996; Harris et al., 1996). The difference in the results for model runs with and without heterogeneous reactions was examined. During the winter time of 1996-2002, the surface areas of NAT and ice are in the order of 10⁹ and 10⁸ m² over the subtropics, respectively, which are 10 or 100 times smaller than those in the polar region. The ClO increases by 0.01-0.27 ppbv (factor of 2-40) and HO₂ increases by 4-26 pptv (factor
of 0.5-2.6) if we considered NAT and ice formation, and the increases of ClO and HO₂ are also smaller than those in the polar region (e.g. Anderson et al., 1991). The ECMWF temperature data over the subtropical western Pacific indicates that the minimum temperatures occur around 100 hPa, with values of 188-192K in winter time during 1996-2002. Such low temperatures and the increase of ClO and HO₂ show that the ozone destruction cycles worked in the model. We estimated that the heterogeneous reaction effect on total ozone was at most 2-3 DU. This magnitude is smaller than the effects of QBO, ENSO, and MSU temperature on the total ozone, but cannot be neglected. Therefore, we conclude that the heterogeneous reaction may occur over the tropics but it has a smaller influence on the variation of the total ozone compared to the influence of the dynamical processes such as QBO. A key question is how much CFCs can be decomposed and converted into Cly in the subtropical lower stratosphere.

4. Conclusions

Using ozone data from EP_TOMS and regression methods developed by Ziemke et al. (1997), the contributions of various time-varying signals to the total ozone over the subtropical western Pacific region were analyzed. The results show that (1) the total ozone variation includes seasonal variability, solar, QBO, ENSO signals, and the MSU temperature variation; (2) the seasonal variability is dominant among them; and (3) the QBO contributed greatly to the unusual ozone lows in the winters of 1996/97, 1998/99, and 2001/02.

The dynamical processes were found to play a major role in the year-to-year ozone variation, which was also simulated by a nudging CTM using ECMWF data, excluding heterogeneous reaction processes over the 45°S-45°N bands. The ozone lows over the subtropical western Pacific were also well simulated. The year-to-year variation in the total ozone is closely related to the position of the ozone minimum, which is largely controlled by the QBO cycle.

However, the discrepancies between the calculated and the observed ozone show that the heterogeneous reaction processes may have some influence on the ozone variation over the subtropical western Pacific. Further comparison of two models, with and without NAT and ice formation, showed that the influence of the heterogeneous reaction processes on the total ozone was less than the influence of the dynamical processes.

Acknowledgments

We thank Mr. Hamada and Mr. Nagasawa of the SX-6 maintenance room in NIES from NEC for improving computing efficiency of the nudging CTM. The GFD-DENNOU library 5.0.1 and the GTOOLS 3.5 were used for Figure 4.

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**Publications and Presentations**

**Original Papers and Reviews:**


**Conference Reports:**


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Hydrological Projection over Asia under the Global Warming with a Regional Climate Model Nested in the CCSR/NIES AGCM

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Abstract

We have developed a regional climate model, NIES-RAMS, and conducted present and future Asian regional climate simulations which were nested in the results of Atmospheric General Circulation Model (AGCM) experiments. The regional climate model could capture the general simulated features of the AGCM and also some regional phenomena such as orographic precipitation, which were not appeared in the outcome of the AGCM simulation, were successfully produced. The increases of annual mean surface runoff and its large fluctuations are projected in a lot of Asian regions. The seasonal mismatch between water demand and water availability, which should be considered in future water resource assessments, is concerned.

Keywords: Hydrological Projection, Global Warming, Asia, Regional Climate Model, CCSR/NIES AGCM

1. Introduction

The impact of hydrological change as a result of global warming caused by anthropogenic emissions of greenhouse gases is a fundamental concern. In order to control water resources in the face of drought, flood, and soil erosion, which frequently present a serious threat to human life and natural ecosystems, risk assessments are being required more frequently by policy-makers.

Though hydrological predictions associated with the global climate change are already being performed mainly through the use of General Circulation Models (GCMs), coarse spatial resolutions and uncertain physical processes limit the available information for water management on a regional scale. Currently, as one of the methods for downscaling the GCM results, dynamical downscaling using a regional climate model (RCM) is in progress.

Higher resolution simulation by a regional climate model is expected to be able to improve regional circulation pattern and mesoscale precipitation such as orographic precipitation resulting from realistic topography.

2. Model and Experiments

We have developed a regional climate model, NIES-RAMS (Emori et al., 2001), which is based on a three-dimensional nonhydrostatic compressible dynamic-equations model (RAMS: Regional Atmospheric Modeling System) developed at Colorado State University (Pielke et al., 1992).

For dynamical downscaling using a RCM which has consistency with the GCM schemes as much as possible, some of the major schemes of the CCSR/NIES AGCM (Numaguti et al.,
1997), Arakawa-Schubert cumulus parameterization (Arakawa and Schubert, 1974), large scale condensation, a radiation scheme based on the two-stream k-distribution method (Nakajima et al., 2000), the Mellor-Yamada level 2.0 turbulence scheme (Mellor and Yamada, 1974, 1982), and the soil and vegetation model (MATSIRO: minimal advanced treatments of surface interaction and runoff) developed by Takata et al. (2003), are incorporated into the RAMS.

Present (1981-1990) and future (2041-2050) Asian regional climates were simulated by the NIES-RAMS nested in the AGCM experiments results (6 hourly) whose horizontal resolution is T42 spectral truncation (about 2.8 degree in each grid). In the RCM experiments, the domain is about 9,600×7,100 km and represented by 161×119 grids with a spatial resolution of 60 km. Nudging timescale of lateral boundary is 600 s and the nudging points from lateral boundary are 10 grids. In the present Asian climate experiment, observed SST and sea-ice distributions for 1981-1990 (GISST: Rayner et al., 1996) were used. In the future experiment, a seasonally varying “warming pattern”, which is derived from a transient climate change experiment with CCSR/NIES coupled ocean-atmosphere climate model, was added to the observed SST.

![Simulated region and topography.](image)

3. Results

3.1 Dynamically Downscaled Present Climate

Dynamical downscaling using the developed regional climate model (RCM) seems to successfully improve regional circulations.

Figure 2 shows observed and simulated present mean precipitation for 10 years (1981-1990) by CPC Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997), the Global Precipitation Climatology Project (GPCP; Huffman et al., 1997), the CCSR/NIES AGCM, and the NIES-RAMS.
RCM improved regional precipitation patterns such as rain band from the Philippine Islands to the East China Sea and Japan which is associated with Baiu fronts, tropical depressions. Orographic precipitation, which can be found obscurely in the AGCM, was also clearly produced in the western part of India and the south-facing slope of the Himalayas by the NIES-RAMS. Not all precipitation patterns improved by the nested RCM. Precipitation of GCM in the tropical ocean is much stronger than the observations and its strongest precipitation zone shifted northward.

Basically, the RCM seems to follow GCM but estimates rather more precipitation. Compared to the observations and GCM, RCM overestimates precipitation in the eastern part of China. That is bias produced by the RCM. By investigating this bias of the NIES-RAMS, it was found that low pressure caused convergence, increased water vapor, and excessive precipitation. The excessive precipitation released much latent heat and it increased air temperature and thickness which caused lower pressure. While lateral boundary conditions are nudged by the GCM, the systematic bias in the RCM was caused by this positive feedback.

![Image](image_url)

**Figure 2** Observed and simulated annual mean precipitation for 1981-1990 by CMAP (upper left), GPCP (upper right), CCSR/NIES AGCM (lower left), and NIES RAMS (lower right).

Figure 3 shows simulated annual mean wind at 850 hPa in present climate. Synoptic circulations, middle-latitude westerlies and monsoon westerlies, were reasonably captured by the RCM. Subtropical high simulated by the GCM relatively weaker than that of ECMWF Re-Analysis (ERA15) (not shown). Simulated weak subtropical high in GCM makes the region more affected by the systematic bias of the RCM. That is, over the eastern part of China, meridional wind arises from convergence caused by the systematic bias. It causes
stronger water vapor transport and precipitation. This might be one of the additional reasons of enhanced precipitation in the region.

![Simulated annual mean wind for 1981-1990 by the CCSR/NIES AGCM (left panel) and the NIES-RAMS (right panel).](image)

3.2 Future Climate Prediction under the Global Warming

About one degree Celsius increase of annual mean 2 m air temperature in the tropical region and about 3-4 degree Celsius in the mid-high latitudinal region are projected from the present climate (1981-1990) to the future (2041-2050) (not shown).

As a result of stronger temperature increase in the higher latitudes, the meridional gradient of air temperature became weak and westerlies caused by thermal wind were weakened (Figure 4). El Niño like differences can be found in the projected results by the AGCM under the global warming. Active convection area over the Western Pacific Ocean shifted to the eastward. As a result, Walker circulations were modulated and the sinking motions were stronger over the Western Pacific Ocean. This caused the negative anomaly of precipitation in the future climate (Figure 4).

Under the global warming, increase of water vapor and higher air temperature over the land than over the ocean were projected. It enhanced a giant land-sea breeze, Asian monsoon circulation. Monsoon westerlies in the Arabian Sea, which is associated with Somali jet, were projected to be stronger and to bring more abundant water vapor to the south of India and the Bay of Bengal (Figure 5). Precipitation was enhanced especially over the mountainous regions, the western part of India and the southern edge of the Tibetan Plateau. Increased vapor flux over the region from the Philippine Islands to the East China Sea which is related with the divergent flow anomaly over the Western Pacific Ocean and increased water vapor was also projected. It increased precipitation in the vicinity of Japan (Figure 4).

As a result of the changes in the synoptic scale flow patterns and precipitation under the global warming, the increase of annual mean surface runoff was projected in a lot of Asian regions (Figure 5). However, the surface runoff changes had regional differences and both the increase and decrease were suggested in summer. It might increase the risk of mismatch between water demand and water availability in the agricultural season.
3.3 Projected Seasonal Hydrological Change under the Global Warming

For accessing the impact of climate change on hydrological cycles in which seasonal cycles such as snow, soil moisture, and evapotranspiration have important role, it is necessary to investigate seasonal changes of hydrological cycles. Six regions were selected arbitrarily as shown in Figure 6.

As one example, heat and water flux changes in Tibet are shown in Figure 7. In Tibet, large part of precipitation is produced by large scale condensation scheme and most of the precipitation consists of snowfall in winter. Convective precipitation can be only found in summer. Snowmelt occurs in early summer and it increases soil moisture and produces more runoff than precipitation. As soil is getting saturated, latent heat flux becomes dominant in the late summer (not shown).

Under the global warming, both temperature and humidity were projected to increase in Tibet (not shown). Snowmelt was projected to occur earlier and increase runoff and soil moisture in spring. Increase of convective precipitation in summer causes the decrease of cloud cover and net downward longwave radiation and the increase of net downward shortwave radiation. Because of increased incoming energy, latent heat was projected to increase a whole year especially in summer. In autumn, rainfall was projected to increase.
instead of decreasing snowfall due to the warming. Because of the decrease of snowfall, increase of precipitation directly recharges soil moisture in autumn.

Figure 6  Selected six regions for investigating the impact of global warming on seasonal hydrological cycles in each Asian region.

Figure 7  Projected seasonal changes of hydrological cycles in Tibet. Changes of heat flux (upper panel) and water flux (lower panel).
4. Summary

We have developed a regional climate model, NIES-RAMS, and conducted present and future Asian regional climate simulations which were nested in the AGCM experiments results. The regional climate model could capture the basic simulated features of the AGCM and also improve some regional phenomena such as orographic precipitation which were not appeared in the outcome of the AGCM simulation.

The increases of annual mean surface runoff and its large fluctuations were projected in a lot of Asian regions. In some regions, the projected seasonal changes of hydrological cycles under the global warming potentially increase the risk of droughts and floods.

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Simulation of the Early Stage of the Last Glacial Period with a Coupled General Circulation Model

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Abstract
A simulation study of the last glacial inception (about 115,000 years before present; 115kyrBP) has been carried out with the Meteorological Research Institute (MRI) coupled General Circulation Model (CGCM). The Earth’s orbital parameters at 115kyrBP and pre-industrial greenhouse gas concentrations were used as the boundary conditions. Longtime integration under the 115kyrBP orbital forcing successfully generated perennial snow cover at Canadian Archipelago which is consistent with geological evidence. This result supports the view that the gradual changes of the ocean and land surface together with the change of the orbital parameters plays the important role for the glacial inception.

Keywords: Last glacial inception, Milankovitch forcing, Coupled GCM

1. Introduction

The last interglacial period (Eemian) began at about 130kyrBP (130,000 years before present) and ended about 117kyrBP (Imbrie et al., 1984). The sea level records show a rapid drop of 50-80 m between 118kaBP and 106kaBP (e.g. Chappell and Shackleton, 1986; Mc-Manus 1994; McCulloch et al., 1999). This lowering is about half of the sea level difference between the last glacial maximum (LGM) and present-day (120 m), and suggests that the equivalent water moved to land as ice masses. Some geological evidence suggests that the Laurentide Ice Sheet initiation started at about 115kyrBP from Baffin Island and Queen Elizabeth Islands (e.g. Andrews, 1982; Andrews et al., 1985; Clark et al., 1993).

The Milankovitch theory teaches us that the summer insolation at northern high latitudes reaches minimum about 116kyrBP for a period of the last 140 k years (a calculation based on Berye, 1979 is shown in Murakami et al., 2002) and cause a suitable condition for the growth of northern ice-sheet. However, the early studies of the last glacial inception with atmosphere only GCMs failed to generate perennial snow cover at appropriate locations under the orbital only forcing (Rind et al., 1989; Oglesby, 1990; Phillipps and Held, 1994; Gallimore and Kutzbach, 1995) In those studies, the present-day SST (sea surface temperature) distribution and landsurface conditions were used with coarse resolution AGCMs. Contrary, some studies using slab ocean coupled GCM and/or vegetation model coupled GCM or experiments using estimated 115 or 116 kyrBP SST succeeded to generate perennial snow covers. (Gallimore and Kutzbach, 1996; de Noblet et al., 1996; Pollard and Thompson, 1997; Dong and Valdes, 1995; Yoshimori et al., 2002). These results suggest the importance of the gradual changes of the surface conditions under the orbital forcing play the important role for the initiation of the glacial period.

In this study, a fully coupled Atmosphere-Ocean GCM is used to investigate the role of the ocean and land surface conditions for the glacial inception.
2. Model and Experimental Design

2.1 Model

The version 2.3 of the Meteorological Research Institute coupled GCM (MRI-CGCM2.3) was used in this study. The atmospheric part of this model is a spectral GCM with T42 resolution and 30 vertical levels (model top at 0.4 hpa). The oceanic part of this model has 2.0 x 2.5 horizontal resolution and 23 vertical levels with model bottom at 5000 m. Flux adjustments for heat and water fluxes at atmosphere-ocean interface were used to reproduce a appropriate present-day climate. A detailed model description is shown in Yukimoto et al. (2001).

2.2 Experimental Design

Two experimental runs were conducted with MRI-CGCM2.3. One is a pre-industrial control run which has present-day orbital parameters and pre-industrial greenhouse gas (GHG) concentrations (see Table 1). The other is a 115kyrBP run which has 115kyrBP orbital parameters and pre-industrial GHG concentrations (Table 1). These two runs have common solar-constant of 1365 W/m² and present-day vegetation map for land-surface conditions. First, we conducted the control run for 100 years from an initial condition taken from a present-day spin-up run. From 101st year, we started the 115kyrBP run with an initial condition taken from the endpoint values of 100 years control run and continued the control run one hundred more years. The 115kyrBP run was conducted about 400 (model) years. The difference of these two runs are only orbital parameters (and initial condition). Thus, we can clearly see the effect of the orbital forcing.

Table 1  Summary of the experimental configurations.

<table>
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<tr>
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<th>Control</th>
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<td>0.01672632</td>
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<tr>
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<td>102.0651</td>
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<td>CH₄</td>
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</tr>
<tr>
<td>N₂O</td>
<td>275 ppbv</td>
<td>275 ppbv</td>
</tr>
<tr>
<td>Solar constant</td>
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<tr>
<td>Vegetation</td>
<td>Present-day</td>
<td>Present-day</td>
</tr>
<tr>
<td>Initial conditions</td>
<td>101st yr of ctl</td>
<td>Present-day</td>
</tr>
</tbody>
</table>
3. Results

Figure 1 shows time series of the global mean surface air temperature (SAT), northern high latitudes (60° N - 90° N) mean SAT and global sea ice volume (each value is annually averaged). After the change in orbital parameters, global mean temperature rises about 0.2° C and it seems to decrease quite gradually. It is still 0.13° C higher than that of control run at the end of 400 years integration (decrease trend is $0.6 \times 10^{-3}$ degree per year). This high global mean SAT is due to insolation difference (see next paragraph). On the other hand, temperature averaged over northern high latitude decreases and sea ice volume increases for first 200 years, and then these seem to be stabilized. Therefore, we compare two runs by taking last 100 years mean of the integrations as those climatology. However, we must note that the temperature of the deep and bottom ocean still changes at the end of each run (the decrease trend of global mean temperature might be related with this fact).

![Figure 1](image.png)

**Figure 1** Time series of (a) global mean surface air temperature (annual mean), (b) high latitude (60°N - 90°N) mean of surface air temperature (annual mean) and (c) global sea ice volume. Green line denotes the 115kyrBP run and black-white line denotes the control run.
Figure 2 shows the differences in zonal mean insolation, SAT and SST as functions of the latitude and month. The 115kyrBP insolation is less in northern summer especially at high latitudes and much in northern winter especially at low latitudes compared to the present-day.

Figure 2  Seasonal variations of zonal mean differences in (a) insolation, (b) surface air temperature and (c) sea surface temperature between 115kyrBP and control (preindustrial). The right hand side column shows annual mean values.
As a result, annual mean 115yrBP insolation is much in low latitude and less in high latitudes and global average of the annual mean insolation is about 0.1% larger than that of present-day. Related with this insolation difference, the differences in zonal mean SAT and SST show similar seasonal variation. It causes cooler summer (Figure 3-a) and relatively warmer winter on northern high latitudes. Figure 3-b shows the difference in annual mean precipitation between 115kyrBP run and the control run. Related to the high annual mean temperature at low and mid latitudes, water vapor in the atmosphere increases in 115kyrBP run and the precipitation increases over all latitudes and seasons compared to the control run. The cool summer air temperature changes some of summer precipitation to snow fall at northern high latitudes in 115kyrBP run (Figure 4). The summer snow fall increases at north and east Siberia, north edge of Scandinavia peninsula, Alaska, Labrador, Keewatin and Canadian Archipelago. However, the regions where the perennial snow cover appears are more restricted. Figure 5 shows the snow cover at August in the control run and its difference between 115kyrBP run and the control run. The Perennial snow cover (mean snow cover near 1 at July and August) appears Greenland, Ellesmere Island, Svalbard Islands and some Arctic islands in the control run. It extends to the north of Baffin Island and Queen Elizabeth Islands in the 115kyrBP run. As mentioned in the introduction, it seems to consist with some geological evidence.

Figure 3 Differences in summer surface air temperature (a) and annual mean total precipitation (b) between 115kyrBP and 0kyrBP.
Figure 4  Snow precipitation at July (monthly mean) in 115kyrBP run (a) and its difference from the control run (b). The values smaller than 0.01 are not plotted.
Figure 5  Snow cover at August (monthly mean) in the control run (a) and difference between 115kyrBP run and the control run (b). The values smaller than 0.01 and negative are not plotted.
To see the difference of the climate between the region where the perennial snow cover appears run and the region where that does not appear in the 115kyrBP run, we plot the seasonal variations of some climatological variables for those regions. Figure 6 shows the seasonal cycle of SAT, snow depth and rain and snow fall at Queen Elizabeth Islands (73° N, 98° W) and north Siberia (Taimyr: 73° N, 100° E). Both points have same latitude and quite similar seasonal variations for each variable in the control run. However, the 115kyrBP seasonal variations at Queen Elizabeth Islands are quite different from those at north Siberia. The summer air temperature is significantly cooler and the winter air temperature is quite warmer than those of north Siberia (or 0kyrBP). The 115kyrBP snow depth at Queen Elizabeth Islands is not cyclic. It means a net positive annual snow accumulation. Contrary, the annual net snow accumulation at north Siberia is zero, although the summer snow fall is not zero and the seasonal snow depth cycle is amplified in 115kyrBP. This difference might be related to the fact that the summer precipitation at Queen Elizabeth Islands is turned into the snow fall in 115kyrBP though much of the summer precipitation at north Siberia still remains rain. However, such differences become clear after several hundred integration and the seasonal variations are quite similar between two regions just after the orbital forcing switch. Therefore, these differences of seasonal cycles between two regions rather suggest the importance of the gradual changes of surface conditions, caused by the Earth's orbital variation, for the glacial inception and such gradual change might be quite sensitive to the surface condition itself.

Figure 6  Seasonal variations of the temperature (top), snow depth (middle) and snow and rain precipitation (bottom) at Queen Elizabeth Islands (left hand column) and north Siberia (right hand column). Green line denotes the 115kyrBP run and black line denotes the control run.
Finally we discuss the meridional heat and water vapor transport at 115kyrBP. Figure 7 shows the difference of total heat transport (black line), oceanic heat transport (green line) and atmospheric water vapor transport (red line) between 115kyrBP and the control run calculated from the radiative fluxes (at the top of atmosphere) and surface heat and freshwater fluxes (at the surface). The total poleward heat transport increases about 0.1 PW in both hemispheres. This is qualitatively consistent with the insolation difference and the surface temperature difference between 115kyrBP and 0kyrBP, and is thought to be a direct consequence of the insolation switch. The oceanic northward heat transport decreases about 0.05 PW at low latitudes. This is related to the slightly weakened Atlantic thermohaline circulation (not shown). However, we must note that the temperature of deep and bottom ocean still varies at the end of 400 years integration. More integrations will be needed to discuss the Atlantic thermohaline circulation in 115kyrBP run. The difference of water vapor transport is somewhat complicated but it basically intensifies the water vapor transport in control run (except southern low latitude). Therefore, it is thought to be a direct result of the enhanced meridional circulation and increased atmospheric water vapor content.

![Heat and WVapor Transport (115ka–0ka)](image)

**Figure 7** Differences in meridional total heat transport (black line), ocean heat transport (green line) and atmospheric water vapor transport (red line) between 115kyrBP and control (preindustrial). The unit of the heat transport is PW, and the unit of water vapor transport is $10^{-3}$ m$^3$/day.

### 4. Summary

A simulation of the last glacial inception (115kyrBP) was conducted with the MRI-CGCM2.3. The Earth’s orbital parameters at 115kaBP and pre-industrial GHG concentrations were used as the boundary conditions. The 115kyrBP insolation slightly warms the annually averaged global mean temperature and increases the atmospheric water vapor content. It also causes the cooler summer in northern high latitudes and relatively warmer winter in northern low latitudes and increases the poleward heat and water vapor transport. These create a suitable condition to initiate the northern high latitudes ice-sheets. The perennial snow cover over Queen Elizabeth Islands and north of Baffin Island were produced in the 115kyrBP simulation run. These result seem to support the view that the gradual changes of the ocean and land surface together with the change of the orbital parameters plays the important role for the glacial inception.
References


Presentations and Publications

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Impact of Different Cloud Modeling Assumptions on Climate Sensitivity in a General Circulation Model

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Abstract
Equilibrium response of surface air temperature to the atmospheric CO₂ doubling is examined using CCSR/NIES/FRCGC atmospheric general circulation model (AGCM) coupled to a slab ocean model to evaluate impact of modifying large scale condensation scheme on the model calculated climate sensitivity. It has been indicated that modification of (1) empirical function which determines the phase of cloud water, and (2) treatment of cloud ice sedimentation resulted in increase of climate sensitivity from 4.0°C to 6.3°C. The obtained results underline the need for observation of cloud water distribution and of more physical treatment of cloud ice processes in the climate model, in order to increase our confidence in the simulated climate change.

Keywords: AGCM, Climate sensitivity, Cloud, Global warming

1. Introduction

The equilibrium response of surface air temperature to a doubling of the atmospheric CO₂ content (“climate sensitivity” hereafter) has become part of the standard evaluation of a general circulation model (GCM), because it is one of the major factors which characterize the response of ocean-atmosphere coupled GCMs to natural/anthropogenic forcings in global warming projections. It was noted that the calculated temperature response varies from model to model as much as 3.0°C in the late 21st century even when the models were driven by similar radiative forcing scenario (IPCC, 2001). Since the difference in the model response is mainly attributed to the differing climate sensitivities and differing rates of oceanic heat uptake in each model, it is desirable that we identify the sources of uncertainty in the model climate sensitivities in order to increase our confidence in the calculated projections.

Many papers have discussed the importance of clouds in determining the sensitivity of climate to changes in greenhouse gases. In a comparison of 19 models forced with sea surface temperature anomalies, Cess et al. (1990) show considerable disparity in both the size and sign of cloud feedback. Although an updated comparison (Cess et al. 1996) shows some convergence between models, uncertainty in cloud processes remains a key issue in understanding the difference in the model sensitivity.

In the present study, we focus on the cloud modelling with large scale condensation process, and discuss how it is related to the climate sensitivity of CCSR/NIES/FRCGC AGCM. Modifying large scale condensation process within the range of uncertainty, two versions of the Atmospheric General Circulation Model (AGCM) are prepared; one with the sensitivity of 4.0°C (“LOW version” hereafter) and the other with 6.3°C (“HIGH version” hereafter). We discuss mechanisms of how the difference in climate sensitivity originated from different assumptions on the cloud treatment. Processes controlling the model climate sensitivity are described, which, we expect, will help develop a strategy for future model improvement to reduce uncertainty in the global warming projections.
2. Model and Experimental Design

2.1 Model

In the following numerical experiments we use CCSR/NIES/FRCGC AGCM coupled with a mixed layer (slab) ocean model. The AGCM is based on the CCSR/NIES model described in Numaguti et al. (1997). The resolution for the AGCM is T42 in horizontal and 20 levels in vertical. Sea surface temperature (SST) and sea ice thickness are calculated using specified ocean heat transport convergence (Q flux) which is required to maintain the present SST and sea ice distribution.

Prognostic cloud water scheme of Le Treut and Li (1991) is used for large scale condensation process, in which total cloud water mass (liquid + ice) is calculated assuming a sub-grid scale distribution of total water. Predicted total cloud water (liquid + ice) is then classified into liquid phase and solid phase using an empirical function which diagnoses the ratio of liquid cloud water mass to total cloud water (liquid + ice) mass by local air temperature (Figure 1). Part of the cloud liquid water and cloud ice are converted to rain and snow, which fall to the ground in a single time step. The rate of this auto-conversion process is set to be more efficient for cloud ice than for cloud liquid water. Cloud ice sedimentation term as described in Senior and Mitchell (1993) is also adopted in the model.

![Graph showing empirical functions for liquid cloud water mass ratio](image)

**Figure 1** Empirical functions which determine the ratio of liquid cloud water mass to total cloud water mass (liquid + ice) by local air temperature. Examples from different AGCMs are displayed.

2.2 Experimental Design

The "LOW version" model is different from "HIGH version" model in the cloud modeling assumptions as described below:

1. The empirical function which determines the ratio of liquid cloud water to total cloud water is different between "LOW version" and "HIGH version" (Figure 1). Cloud ice is allowed to exist in higher temperature regime (lower latitude/altitude) in "LOW version" compared with in "HIGH version". Each function adopted by different models as shown in Figure 1 is based on different observations, suggesting that the functions for both "LOW" and "HIGH" versions are within the range of uncertainty.
2. Melted cloud ice after sedimentation is classified into rain in “LOW version”, while it is classified into cloud liquid water suspended in the atmosphere in “HIGH version”. Therefore, less cloud water tends to exist near the melting temperature in “LOW” model compared to in “HIGH” model.

The “LOW” and “HIGH version” models are integrated with pre-industrial (year 1850) condition for 45 years in the control experiments. Both models show similar capability in reproducing the present climatology when evaluated with the Climate Prediction Index described in Murphy et al. (2004). The two models are then integrated with doubled CO$_2$ condition for 45 years. Averages of the last five years of integrations are used to compare the equilibrium response to CO$_2$ doubling between “LOW” and “HIGH” models.

3. Results

Time series of global annual mean 2 m temperature are displayed in Figure 2, showing that equilibrium response to CO$_2$ doubling amounts to 6.3°C in “HIGH version” and 4.0°C in “LOW version”. Temperatures in the control experiments of both models remain relatively stable and show little drift during the integrations. The results imply that model climate sensitivity is highly dependent on the treatment of liquid/ice cloud formation process, which is consistent with the findings of Li and Le Treut (1992).

![Figure 2](image.png)

**Figure 2** Time series of global annual mean 2 m temperature in control experiment (lower two curves) and doubled CO$_2$ experiment (upper two curves). Black curves and red curves correspond to the results of the “LOW version” and the “HIGH version”, respectively.

To understand how the temperature response to CO$_2$ increase is affected by cloud change, response of cloud radiative forcing is examined first in Figure 3(a)-(c). In both models (“LOW” and “HIGH”), net cloud radiative forcing (shortwave plus longwave contribution) decreases in low to middle latitudes and increases in high latitudes, suggesting that cloud response to the CO$_2$ increase enhances the warming in low-middle latitudes while it suppresses the warming in high latitudes. This feature is mainly due to the response of shortwave radiative forcing, partly compensated by the contribution of longwave components with opposite sign. The response of cloud radiative forcing as described above reflects the response of cloud water distribution to CO$_2$ increase as shown in Figure 3(d)-(f). Cloud water increase is evident in the region enclosed by two isotherms, corresponding to where ice cloud and liquid cloud co-exist. This increase in cloud water suppresses the warming by increasing the shortwave radiative forcing in high latitudes. The cloud water increase can be explained...
by the fact that, as temperature warms up, ice cloud is replaced by liquid cloud which has less efficient autoconversion rate than ice cloud, leading to decrease of sink term in the cloud water tendency equation (Mitchell et al., 1989). In low latitudes, cloud water increases in boundary layer and middle troposphere while it decreases in between the two layers (sigma=0.7-0.8). The radiative forcing change which enhances the warming in low latitudes can be interpreted as the vertically integrated impact of these cloud water change, dominated by the cloud water decrease in the middle layer (sigma=0.7-0.8). Impact of cloud on surface downward shortwave radiation ("S-SCRF" hereafter) is calculated as another diagnostic similar to the shortwave cloud radiative forcing. Vertical profile of contribution to the S-SCRF (not shown) indicates that cloud water decrease in the middle layer is responsible for the increase in surface downward shortwave radiation, which support the interpretation presented above.

Difference of cloud response to CO₂ increase between "LOW" and "HIGH" versions are discussed next, to clarify how climate sensitivity is affected by the modifications in cloud scheme. If we compare the net cloud radiative forcing response between "LOW" and "HIGH" versions, we notice that decrease in low latitudes which enhances the warming is restricted to smaller regions in "LOW" version compared to in "HIGH" version. Considerable reduction of net (and shortwave) radiative forcing can be seen in southern middle latitudes (50-60S) in Figure 3(b), while similar characteristics cannot be found in Figure 3(a). The difference of radiative forcing response between "LOW" and "HIGH" versions (Figure 3c) amounts to as much as 10 W m⁻² in southern middle latitudes, suggesting that cloud response of "LOW" version suppresses the warming compared to the response of "HIGH" version. This difference in radiative forcing response between "LOW" and "HIGH" version is consistent with the cloud water response illustrated in Figure 3(d)-(f). Cloud water increase related to liquid-ice phase transition is located in lower latitudes/altitudes in "LOW" version than in "HIGH" version. Accordingly, cloud water decrease which causes the radiative forcing reduction in low latitudes is restricted to smaller regions in "LOW" version than in "HIGH" version. The difference of cloud water response between "LOW" and "HIGH" versions is most pronounced in southern middle latitudes (Figure 3f), which is consistent with the marked difference of radiative forcing response shown in Figure 3(c).

The difference of cloud water response as described above can be attributed to the difference of cloud modelling assumptions between the two models. First, the difference of the empirical function as illustrated in Figure 1 allows mixed phase clouds to exist in lower latitudes/altitudes in "LOW" version than in "HIGH" version. Therefore, in response to CO₂ increase, ice cloud is replaced by liquid cloud, leading to cloud water increase in lower latitudes/altitudes in "LOW" version. Second, the different treatment of melted cloud ice results in different cloud water content in the control experiments. We expect that less cloud water in the "LOW" control experiment also contributes to the lower climate sensitivity by moderating the impact of cloud water decrease on planetary albedo, although further discussion is needed to describe the mechanism of how the sensitivity is related to the cloud water content.
Figure 3  Response of annual zonal mean cloud radiative forcing (a)-(c) and response of annual zonal mean cloud water (d)-(f) to doubling of the atmospheric CO$_2$ concentration. Blue curves and red curves in (a)-(c) indicate the response of longwave and shortwave component, respectively. Black curves show the response of net cloud radiative forcing (longwave plus shortwave). (a) and (d) show the results of “LOW version”, (b) and (e) the results of “HIGH version”, while (c) and (f) correspond to the difference of the two results; “LOW version” minus “HIGH version”. Radiative forcing is defined positive upward. Isotherms of control experiments (0°C, -5°C) and doubled CO$_2$ experiments (-15°C, -25°C) are also shown in the panels (d) and (e).

4. Summary

Equilibrium response of surface air temperature to the atmospheric CO$_2$ doubling is examined using CCSR/NIES/FRCGC AGCM coupled to a slab ocean model, in order to discuss the impact of modifying large scale condensation scheme on the model calculated climate sensitivity. The obtained results indicate that modification of (1) empirical function which determines the phase of cloud water, and (2) treatment of cloud ice melted due to sedimentation, within the range of uncertainty, resulted in increase of climate sensitivity from 4.0°C to as high as 6.3°C. Definition of the empirical function, (1), affects the climate sensitivity through changing the latitude and height at which cloud water content responds to CO$_2$ increase, which are closely related to the response of cloud radiative forcing. Particularly, response of cloud water over the Southern Ocean plays a crucial role in controlling the model sensitivity.

Results of the present study support the argument that climate sensitivity is highly dependent on the control cloud ice distribution and its response to CO$_2$ increase. Moreover, it should be emphasized that latitudinal distribution of cloud water response to CO$_2$ increase is one of the key factors which control the model sensitivity. The obtained results underline the importance of both (a) observing cloud water distribution and (b) more physical treatment of
cloud ice generation processes in CCSR/NIES/FRCGC model, in order to increase our confidence in the simulated climate change.

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Radiative Convective Equilibrium Calculations with a Cloud Resolving Model toward Global Cloud Resolving Experiments

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Abstract
A set of radiative convective equilibrium experiments are proposed to compared to results with a global cloud resolving model which is being developed by authors. It is thought that radiative convective equilibrium calculations corresponds to climate states under a uniform tropics condition, but it lacks effects of large scale motions represented as the Hadley and Walker circulations. Comparison between radiative convective equilibrium experiments and the aqua planet experiments will reveal the interaction between mid-latitude circulation and low-latitude circulation, that is the interaction between individual clouds in the tropics and the large scale circulation. This study reports one aspect of the radiative-convective experiments calculated with a three-dimensional cloud resolving model with periodic boundaries in the Cartesian coordinate. Statistics of vertical velocity are examined with emphasis on the dependency on the domain size and the grid interval. If the grid number is the same, the statistics are very similar as long as the grid interval is less than 4km. It is argued that the grid number should be at least $100 \times 100$ in order that multiple clouds coexist within the domain.

Keywords: Radiative Convective Equilibrium, Cloud Resolving Model, Global Cloud Resolving Model, Aqua Planet Experiment

1. Introduction

One of the most ambiguous factors of climate modeling comes from cloud parameterizations. It is thought that this issue will be resolved by using cloud resolving models with which cloud motions are directly resolved. Recently, a global cloud resolving model is being developed (Tomita and Satoh, 2004); this model is based on the nonhydrostatic equations and the icosahedral grids and is aimed to be used for climate study. Although global cloud resolving experiments can be executed with a very special computer (i.e. the Earth Simulator), there are many things to be examined before the new model is used as a reliable climate model. In order to evaluate results with global cloud resolving experiments, we must compare them with results with other models or other experimental configuration using various computer resources.

The aqua planet experiment (APE; Neale and Hoskins, 2001) is suitable for the first step to the global cloud resolving modeling: APE is a series of experiments by giving a sea surface temperature distribution on the globe. In order to study climate states in the tropics, in particular, it is useful to compare results of APE with those with simplified experiments under a uniform tropical condition. The latter type of experiments are known as radiative convective equilibrium experiments, and are performed by many researchers. In this study, we report one aspect of the radiative convective equilibrium experiments in order to understand results of APE with the global cloud resolving model and other general circulation models. This
comparison will lead to better understanding of interaction between mid-latitude circulation and low-latitude circulation.

2. Radiative convective equilibrium experiments

Radiative convective equilibrium experiments (RCE) are performed with a different model configuration. First, we use a Cartesian coordinate nonhydrostatic model (Satoh 2003) with double periodic boundary conditions. The domain size and the grid interval are one of the key factors of the experiments. Tompkins and Craig (1998) chose the 100km×100km domain size with $\Delta x = 2$km grid interval on the basis of the estimation of vertical velocity. The case with this domain size and the grid interval is referred to as the control one. We change the domain size and the grid interval to study statistical properties of climate states. Satoh et al. (2004) have further extended the domain size to cover an equatorial belt with 40,000km×100km by using JMA-NHM. This experiment will be directly compared with a global cloud resolving aqua planet experiment, and will reveal the importance of the interaction between the large-scale motion and individual cloud motions. A global cloud resolving model developed by Tomita and Satoh (2004) can also be used to study RCE by specifying a uniform sea surface temperature assuming everywhere tropics condition. The radius of the earth is arbitrarily chosen depending on computer resources. The radius of the earth is set as $R = 100$km, 200km and 400km, for instance. These can be called “small planet experiments”. In the case of $R = 400$km, the surface area of the planet is $4\pi R^2 = (1,418km)^2$, which is sufficiently large to study statistical properties of tropics. It is informative to run RCE on different models. We show in this report the results of a Cartesian coordinate model by Satoh (2003).

Figure1 shows statistics of vertical velocity in different domain size and grid interval: 1. 100km×100km with $\Delta x = 2$km, 2. 200km×200km with $\Delta x = 2$km, 3. 200km×200km with $\Delta x = 4$km, and 4. 500km×500km with $\Delta x = 10$km. Right panels show time sequences of maximum and minimum vertical velocity within the domain. Left panels show probability distribution functions (pdf) of maximum vertical velocity, which is calculated from maximum vertical velocity at each time step. In the case of the control experiment 1, maximum vertical velocity reaches 15-20m/s less frequently compared to case 2. The average maximum vertical velocity is most frequent at 3m/s in that of case 1, while it is most frequent at 10m/s in case 2. Cases 3 and 4 also have a peak around 3-5m/s in pdfs, which show a distribution similar to case 1. These three cases have the same grid number 50×50 within the domain. The maximum vertical velocity of case 4 is at 10m/s, which is very small compared to other three cases. This case has a very coarse grid interval $\Delta x = 10$km, which is not sufficient to resolve cloud updrafts. In case 2, pdf is most broad, and the maximum vertical velocity reaches about 20m/s. It is expected that if the domain size is sufficiently large, strong cloud updrafts may coexist and the strong maximum vertical velocity will be observed somewhere in the domain, so that the pdf will have a peak at strong vertical velocity. From these results, the grid number should be more than 100×100 in order that multiple cumulus clouds coexist within the domain. This conclusion is different from Tompkins and Craig (1998), in which they used 100km×100km domain size with $\Delta x = 2$km in order to obtain sufficient speed of vertical velocity. Although the maximum vertical velocity is at the similar level for cases 1, 2, and 3 (the grid interval is less than 4km), the larger grid number 100×100 is more appropriate since more than one cloud coexist within the domain.
Figure 1  Statistics of vertical velocity of radiative convective experiments. Left: probability distribution functions of maximum vertical velocity within the domain. Right: time sequences of maximum and minimum vertical velocities in the domain. From top to bottom, 1. 100km × 100km with ∆x = 2km, 2. 200km × 200km with ∆x = 2km, 3. 200km × 200km with ∆x = 4km, and 4. 500km × 500km with ∆x = 10km.
3. Summary

We proposed a set of radiative convective equilibrium experiments in order to compare results with a global cloud resolving model. Comparison between the radiative convective experiments and an aqua planet experiment with global models will reveal the interaction between the mid-latitude circulation and the low-latitude circulation. This is also the interaction between the individual cloud motions and the large-scale circulation represented as the Hadley and Walker circulation. We studied the statistics of the maximum vertical velocity of the radiative convective experiments, particularly on the dependency on the domain size and the grid interval. It is found that the grid number should be at least 100 × 100 in order that multiple clouds coexist within the computational domain.

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Publications and Presentations

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Ozone QBO Simulated by the Stratospheric Chemical Transport Model of Meteorological Research Institute (MJ98-CTM)

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Abstract
The ozone quasi-biennial oscillation (QBO) in the equatorial stratosphere is realistically reproduced with the 3-dimensional chemical transport model of Meteorological Research Institute (MJ98-CTM) and the properties of the simulated QBO are investigated through the comparison with observed one. MJ98-CTM is comprised of a general circulation model and a chemical transport module with full chemistry for the middle atmosphere and ozone is treated interactively. T42L68 version of MJ98-CTM simulated the zonal wind QBO, though its amplitude is weaker, of about 31 months period, being very close to observed average value of 28 months. The QBO structure was also realistically captured as proved that the temperature component preceded the wind component by about 6 months. The simulated ozone QBO shows amplitude of about 0.4 ppmv in the middle stratosphere and it precedes the wind QBO by about 4 months at 30 hPa, similarly to observations.

Keywords: quasi-biennial oscillation (QBO), ozone QBO, Interactive ozone, Equatorial stratosphere, Chemical transport model

1. Introduction

The zonal wind in the equatorial middle and lower stratosphere has been well known to alternate its direction with the period of approximate two years from the confirmed findings in 1961 (Reed et al., 1961; Veryard and Ebdon, 1961). This phenomenon called the quasi-biennial oscillation (QBO), named by Angell and Korshover (1964), originates above 30 km and gradually descends at a rate of about 1 km/month with the amplitude of about 20 m/s over the equator. The amplitude of the QBO is latitudinally symmetric and also meridionally symmetric with a half width of about 12° latitude. The QBO have gathered much scientific interest partly because of its peculiar and broad periodicity ranging from under 2 to about 3 years centered on 28 months, being in distinct contrast to the semiannual and annual cycles of solar forcing on the atmosphere and earth.

The mechanism of the QBO is first provided by Lindzen and Holton (1968), who demonstrated that the vertically propagating gravity waves of a broad spectrum could make the alternation of the mean flow direction through the wave-mean flow interaction occurring immediately below the critical layer. After that, also equatorial Kelvin waves and Rossby-gravity waves are proved to contribute to the driving of the QBO through radiative damping (Holton and Lindzen, 1972).

The effects of the QBO, though its amplitude is confined in the tropics, cover the whole winter stratosphere through the modulation of planetary wave propagation from the troposphere as demonstrated by Holton and Tan (1980). The modulation, in turn, changes wind and temperature in the extratropics including the polar region. The effects can be more clearly seen in the winter stratospheric circulation when stratified according to the solar
11-year cycle (Labitzke, 1987; Naito and Hirota, 1997).

The QBO affects the distribution of atmospheric constituents such as water vapor, chemical species and aerosols through the secondary meridional circulation, so that the QBO component clearly appears in ozone (Funk and Garnham, 1962; Randel and Cobb, 1994; Hasebe, 1994) and it is called the ozone QBO. In contrast to the distinct signal of the ozone QBO in observations, mechanistic chemical models and nudging chemical transport models (Lin and London, 1986; Gray and Pyle, 1989; Zhou et al., 2003), reproducing the ozone QBO with 3-dimensional chemical model was difficult until recent, because general circulation models (GCMs) can barely simulate the wind QBO since the first success of 1.5-year QBO by Takahashi (1996). A GCM capable of simulating the wind QBO, even though far from the realistic one, is thus indispensable for reproducing the ozone QBO.

Nagashima et al. (1998) first reproduced the ozone 1.5-year QBO by interactively coupling the ozone produced through the Chapman reactions with the GCM of Takahashi (1996). Also in this study, we simulated the ozone QBO with the 3-dimensional chemical transport model (CTM) of Meteorological Research Institute, MJ98-CTM, by interactively coupling MJ98 GCM (Shibata et al., 1999) with a chemical transport module, in which a full chemistry for the stratosphere is included (Shibata et al., 2005). T42L68 version of MJ98-CTM successfully reproduced the zonal wind QBO, though its amplitude is weaker, of about 31 months period, being very close to observed average value of 28 months. The QBO structure was also realistically captured as proved that the temperature component preceded the wind component by about 6 months. The simulated ozone QBO shows amplitude of about 0.4 ppmv in the middle stratosphere and it precedes the wind QBO by about 4 months at 30 hPa, similarly to observations.

2. Model and Experiments

To the standard version of MJ98, T42L45 (Shibata et al., 1999), three modifications are made for vertical resolution, gravity wave drag, and horizontal diffusion under the constraint that the effect on the climate in the troposphere be smallest. (I) The number of layers is so increased from 45 to 68 that the vertical grid spacing becomes 500 m in the lower and middle stratosphere from 100 to 10 hPa with gradational layers at lower and upper boundary of this smaller spacing layer, as depicted in Figure 1. (II) In stead of the Rayleigh friction, Hines (1997) gravity wave drag scheme is introduced, wherein a homogeneous source of 1.5 m/s and tropically enhanced source of Gaussian form with 1.0 m/s peak value at the equator are launched at the model lowest level, as illustrated in Figure 2. (III) Horizontal diffusion is so weakened that the e-folding time at the maximum wavenumber 42 is gradually increased from 18 hours at 150 hPa to 180 hours at 100 hPa, above which the same 180 hours is used.

The chemical module is based on the family method and contains major stratospheric species, i.e., 34 long-lived species including 7 families and 15 short-lived species with 79 gas phase reactions and 34 photochemical reactions. Two types (I and II) polar stratospheric clouds (PSCs) and sulfate aerosols are included with six and three heterogeneous reactions on PSCs and sulfate aerosols, respectively. The transport process uses a semi-Lagrangian scheme, which is of flux form in the vertical and of ordinal type in the horizontal. Transport equation is solved so as to be compatible with the continuity equation and thereby the mass of chemical species is well conserved. The detail is described in Shibata et al. (2005).
Ozone is interactively treated between GCM and CTM, and this version T42L68 is integrated from a certain initial condition for the atmosphere under climatological values for sea surface temperature, the greenhouse gases, and sources gases for chemical species. Analysis is made only for nearly stabilized last 15 years.

3. Dynamical Aspects of Simulated QBO

Figure 3 shows the time-pressure cross section of zonal-mean zonal wind averaged between 5°S and 5°N in the stratosphere from 100 to 5 hPa for 15 years. It is evident that the model reproduces realistically the QBO, though the amplitude is somewhat smaller, in particular the westerly phase, than observed one. When the climatological seasonal cycle is subtracted, the westerly phase, though very weak, descends down to 100 hPa level and the amplitudes of both phases become of similar magnitude \( \sim 15 \) m/s at 20 hPa, being about 3/4 of the observed one (Huesmann and Hitchman, 2001). Along with the zonal wind amplitude, the temperature amplitude (\( \sim 2K \) at 20 hPa) is also weaker (not shown) than the observed one (Randel et al., 1999).

The QBO period is evaluated from the auto-correlation of the zonal-mean zonal wind, in which the climatological seasonal cycle is removed. The auto-correlation in Figure 4 shows that the simulated QBO period is about 31 months, which is very similar to the observed period of about 28 months. However, there is so prominent discrepancy between the simulation and observation that the simulated QBO signal extends up to the upper stratosphere, while the observed QBO signal is confined below about 10 hPa, above which the QBO signal is fading with altitude.
Figure 3  Time-pressure cross section of zonal-mean zonal wind in the equatorial stratosphere for 15 years. Zonal wind is averaged between 5°S and 5°N. Black-solid and dashed-red contours indicate westerly and easterly winds. Contour interval is 5 m/s. Note that the wind is raw data, i.e., climatological seasonal cycle is not subtracted.

Figure 4  Vertical profiles of auto-correlation coefficients of the zonal-mean zonal wind averaged between 5°S and 5°N for the model (left) and observation (right) based on 20 years of ERA-40 Reanalysis from 1981 to 2001.

The phase relationship between the QBOs in temperature and zonal wind is shown in Figure 5. The temperature signal precedes the zonal wind signal and the lag slightly tilts toward negative values with altitude in the observed data (Niwano and Shiotani, 2001). The lag attains 6 months, about a quarter cycle, in the middle stratosphere, which relation can be explained by the conventional view, in which the secondary meridional circulation is required to maintain the QBO temperature structure (Hasebe, 1994). The model qualitatively reproduces this phase relationship although the lag-correlation coefficient is underestimated below 50 hPa and overestimated above 10 hPa.
Figure 5 Vertical profiles of lag-correlation coefficients between the temperature and zonal wind for the model (left) and observation (right). Negative lags represent that the temperature QBO precedes the zonal wind QBO.

The forcing of the zonal wind QBO is brought about by the momentum deposition of upward-propagating waves excited in the troposphere. First, Lindzen and Holton (1968) provide a theory due to the gravity waves, which break at the critical layer, and later Holton and Lindzen (1972) updated the theory exclusively favoring the Kelvin waves and Rossby-gravity waves, which deposit their momentum below the critical layer through the radiative damping. However, the gravity waves are now again believed to be the dominant forcing (Dunkerton, 1997), because the Kelvin waves and Rossby-gravity waves are insufficient to force the zonal wind QBO (Gray and Pyle, 1989; Alexander and Holton, 1997; Sato and Dunkerton, 1997).

In the model, there are two types of forcing. One comes from the resolved waves, which are determined by the model resolution and the other stems from small scale gravity waves represented with the parameterization scheme of Hines (1997). The former forcing is calculated by the Eliassen–Palm (EP) flux divergence (EPD), while the latter forcing is directly obtained by gravity wave drag (GWD). In this paper, EP-flux is made of anomalies from daily mean fields. A comparison of the both forcings (Figure 6) revealed that the positive EPD (eastward forcing) occurs during the easterly phase and westerly shear, i.e., the latter half period in the easterly phase, while the positive GWD works dominantly during the westerly phase and westerly shear, i.e., the former half period in the westerly phase. In other words, during the positive forcing the former half acceleration is made by EPD in the easterly phase, following which the latter half is conducted by GWD in the westerly phase. Similarly, during the negative forcing (westward forcing), the former half acceleration is due to EPD in the westerly phase, while the latter half is due to GWD in the easterly phase. The duration of acceleration is shorter during the westerly shear than during the easterly shear both for GWD and EPD, being in compatible with the intensity of the wind shear.

The lag-correlation between the forcings and zonal wind shows more clearly the temporal relation between them as shown in Figure 7, wherein the lag-correlation for EPD calculated from ERA40 is also drawn. The forcing due to the resolved waves precedes the zonal wind by about a quarter cycle, while the phase lag of the forcing due to the gravity waves is very small.
4. Chemical Aspects of Simulated QBO

Figure 8 depicts the time-pressure cross section of deseasonalized zonal-mean ozone averaged between 5°S and 5°N for 15 years, along with the zero-contours of zonal wind. The ozone QBO is apparently reproduced except for the initial cycle of the zonal wind QBO. This
is probably because the abundances of some chemical species, particularly NO\textsubscript{2}, are not yet stabilized in this period. The simulated ozone QBO shows amplitude of about 0.4 ppmv, and the phase below 15 hPa is anti-correlated with that above, being in good agreement with observations (Hasebe, 1994; Randel and Wu, 1996). In the simulated ozone QBO below 15 hPa the positive ozone phase appears sharply in the strong westerly shear, while the negative ozone phase appears broadly in the midst of the easterly phase, being very similar pattern in w* (not shown). The lag-correlation between ozone and w* quantifies this phase relation that they are in phase below 15 hPa and anti-correlated above (not shown).

![Figure 8](image)

**Figure 8** Time-pressure cross section of zonal-mean ozone anomaly averaged between 5°S and 5°N for 15 years. Black-solid and dashed-red contours indicate positive and negative values and blue lines show the zero-contour of zonal wind. Contour interval is 0.1 ppmv.

The phase relation between ozone and zonal wind is illustrated in Figure 9. The lag-correlation coefficient attains maximum and minimum in the middle stratosphere in 20-30 hPa: a maximum of about 0.6 at -4 months lag and a minimum of about -0.6 at 11 months below 15 hPa, above which the coefficients are oppositely signed. These features are very similar to those based on observations (Hasebe, 1994; Niwano and Shiotani, 2001), wherein the precedent phase lag is referred to as about several months or a quarter cycle.

![Figure 9](image)

**Figure 9** Vertical profile of lag-correlation coefficient between ozone and zonal wind. Negative lag means that ozone precedes zonal wind.
5. Discussion and Summary

As demonstrated above, the ozone QBO precedes the zonal wind QBO by about a quarter cycle in the middle stratosphere. Hasebe (1994) ascribed this phase relationship not to the ozone chemistry but to the ozone solar heating from numerical experiments with the 1-dimensional model of simplified Lin and London (1986). However, Huang (1996) denied the ozone solar heating effect on the quarter cycle lag between the QBOs in ozone and zonal wind. Huang (1996) demonstrated that the secondary meridional circulation associated with the zonal wind QBO could solely make the ozone QBO with a quarter cycle precedent phase from the zonal wind QBO. Also we obtained the same results through non-interactive ozone simulations (not shown).

However, these results deny only the effect of the ozone solar heating on the phase relation in QBO components, and thereby do not neglect its effect on the QBO itself. The comparison of interactive and non-interactive ozone simulations demonstrates that the ozone solar heating does elongate the QBO period noticeably (not shown).

For a passive tracer such as N$_2$O below the middle stratosphere, the phase relation can be described with the conventional view and the passive tracer QBO would be in phase with the zonal wind QBO. This phase relation is certainly reproduced in the model as drawn in Figure 10, wherein the phase lag is almost zero in the lower stratosphere, and still as small as about 2 months at 10 hPa without no phase reversal as in ozone throughout the middle atmosphere. Then, the difference of the phase relations with zonal wind between ozone and N$_2$O is considered to come from chemical effect, because they are transported through the same atmospheric circulation.

![Log-correlation N2O vs. U Hines1 (Int)](image)

Figure 10  Same as Figure 9 except for N$_2$O.

T42L68 version of MJ98-CTM is so made from the standard T42L45 version to simulate the QBO. There are three modifications. (I) The number of layers is so increased from 45 to 68 that the vertical grid spacing becomes 500 m in the lower and middle stratosphere from 100 to 10 hPa with gradational layers at lower and upper boundary of this smaller spacing layer. (II) Instead of the Rayleigh friction, Hines (1997) gravity wave drag scheme is introduced, wherein a homogeneous source of 1.5 m/s and tropically enhanced source of Gaussian form with 1.0 m/s peak value at the equator are launched at the model lowest level,
as illustrated. (III) Horizontal diffusion is so weakened that the e-folding time at the maximum wavenumber 42 is gradually increased from 18 hours at 150 hPa to 180 hours at 100 hPa, above which the same 180 hours is used. T42L68 reproduced the zonal wind QBO, though its amplitude is weaker, of about 31 months period, being very close to observed average value of 28 months. The QBO structure was also realistically captured as proved that the temperature component preceded the wind component about 6 months. Simulated ozone QBO shows amplitude of about 0.4 ppmv in the middle stratosphere and it precedes the wind QBO by about 4 months at 30 hPa, similarly to observations.

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**Presentations and Publications**


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Study of the Estimate of New Climate Change Scenarios
Based on New Emission Scenarios
-IPCC AR4 Experiments-

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Abstract
Using the SRES scenarios (A1B, A1T, A1FI, A2, B1, B2) with some extension to stabilization scenarios proposed for the IPCC Fourth Assessment Report (AR4), we have performed the AR4 scenario experiments. The data sets obtained from the experiments have been submitted to the IPCC Model Output which was established to support the Working Group I component of the AR4 at the Program for Climate Model Diagnosis and Intercomparison (PCMDI). Annual and global mean surface air temperature rises by about 2.4, 2.1, 3.2, 2.7, 1.7 and 2.0 degrees centigrade for A1B, A1T, A1FI, A2, B1 and B2 scenarios, respectively, between 1961-1990 and 2071-2100. Geographical response patterns in surface temperature, precipitation and sea level fields are rather similar among the scenarios although their global mean responses are different.

Keywords: IPCC, SRES scenario, Fourth Assessment Report (AR4)

1. Introduction

The purpose of this study is to perform new scenario experiments which are recommended by the Asian-Pacific Integrated Model (AIM) group at the National Institute for Environmental Studies (NIES). Meanwhile the Intergovernmental Panel on Climate Change (IPCC) requested modeling groups around the world to perform an unprecedented set of coordinated the 20th and the 21st century climate change experiments, in addition to commitment experiments extending to the 22nd century. We decided to perform the requested experiments by IPCC instead of original planed ones, because one of the scenarios selected for the use of ensemble runs was the A1B scenario which had been recommended by the AIM group.

The scenarios are originally provided as emission scenarios and concentration scenarios are additionally provided only for well-mixed greenhouse gases. However, the present climate models need atmospheric concentration scenarios to perform the experiments because a chemical transport model (CTM) is usually not incorporated in them. Therefore we firstly calculated the sulfate aerosol distribution off-line by using a CTM developed at the Meteorological Research Institute (MRI). Then we have performed the whole set of the scenario experiments. Now we have submitted the data sets obtained from the experiments to the IPCC Model Output which was established to support the Working Group I component of the AR4 at the Program for Climate Model Diagnosis and Intercomparison (PCMDI). This paper reports results from the new scenario experiments for several basic physical variables.
2. Models

The models used for the present experiments were described in a previous report (Uchiyama et al., 2003). We repeat the essential parts of the models here for the reference.

In order to calculate geographical sulfate aerosol distributions from sulfur dioxide emission scenarios, we used a three-dimensional aerosol CTM, called the Model of Aerosol Species IN the Global AtmosphereRe (MASINGAR: Tanaka et al, 2003), developed at the MRI. The MASINGAR is an on-line CTM for atmospheric aerosol species and is coupled with an atmospheric general circulation model and a four-dimensional data assimilation system. The atmospheric part of the model is a spectral atmospheric GCM, MRI/JMA98 GCM (Shibata et al, 1999), developed at the MRI and the Japan Meteorological Agency (JMA). The model accounts for advective transport, subgrid-scale eddy diffusive and convective transport. The advective transport for trace species in the model is performed using a three-dimensional semi-Lagrangian advection scheme. Parameterization of convective transport is based on the convective mass flux derived by the Arakawa-Schubert scheme.

The MRI-CGCM 2.3 (Yukimoto et al, 2002) is used for global warming experiments. The atmospheric part of the model has a horizontal resolution of T42, and 30-layer sigma-pressure hybrid coordinate. The ocean part of the model has a horizontal grid spacing of 2.5 degrees in longitude and 2 degrees in latitude, and 23-layer vertical level. Near the equator, the meridional grid spacing is set to 0.5 degrees to resolve equatorial Kelvin and Rossby waves. Deep moist convection with a prognostic Arakawa-Schubert scheme and vertical diffusion with a Mellor-Yamada level-2 turbulence closure scheme are used. The sea ice model predicts compactness and thickness. In radiation scheme, CO$_2$, H$_2$O, O$_3$, CH$_4$ and N$_2$O are treated as the greenhouse effect gases directly, and the direct effect of sulfate aerosol is explicitly treated. The magnitude of the indirect effect is highly uncertain due to insufficient understanding of chemistry and microphysics in cloud. Therefore we did not explicitly include the indirect effect in the present calculations.

3. AR4 Scenarios

We have performed the AR4 experiments with the MRI-CGM2. Below is a list of runs for the AR4 including three other scenarios of the Special Report on Emission Scenario (SRES):

1. 20th century simulation to year 2000, then fix all concentrations at year 2000 values and run to 2100 (CO$_2$ $\sim$ 360ppm) (20C3M, Commit)
2. 21st century simulation with SRES A1B to 2100, then fix all concentrations at year 2100 values to 2200 (CO$_2$ $\sim$ 720ppm) (SRESA1B)
3. 21st century simulation with SRES B1 to 2100, then fix all concentrations at year 2100 values to 2200 (CO$_2$ $\sim$ 550ppm) (SRESB1)
4. 21st century simulation with SRES A2 to 2100 (SRESA2)
5. 1% CO$_2$ run to year 80 where CO$_2$ doubles at year 70 with corresponding control run (1%to2x, PDrntl)
6. 100 year (minimum) control run including same time period as in 1 above (PDrntl)
7. 2xCO$_2$ equilibrium with atmosphere-slab ocean
8. Extend one A1B and B1 simulation to 2300 (SRESA1B, SRESB1)
9. 1% CO$_2$ run to quadrupling with an additional 150 years with CO$_2$ fixed at 4xCO$_2$ (1%to4x)
10. 1% CO₂ run to doubling with an additional 150 years with CO₂ fixed at 2xCO₂ (1%to2x)

11. Atmospheric Model Intercomparison Project (AMIP) run

In order to estimate the uncertainty due to natural variability, we have made 5-member ensemble runs for some scenarios (2OC3M, SRESA1B, SRESB1 and SRESA2). Single member runs are also conducted for A1T, A1FI and B2 scenarios.

4. Results

4.1 Global Mean Response

Figure 1(a) shows temporal changes in global and annual mean surface air temperature for all 6 SRES scenarios. Each ensemble member is also shown for A1B, A2 and B1 scenarios. Surface air temperature rises associated with the increase in GHGs in each scenario, with some annual variability and some dispersion among ensemble members. Surface air temperature rises about 2.4, 2.1, 3.2, 2.7, 1.7 and 2.0 degrees centigrade for A1B, A1T, A1FI, A2, B1 and B2 scenarios, respectively, between 1961-1990 and 2071-2100.

Figure 1(b) shows temporal changes in global and annual mean of total precipitation. Precipitation increases in proportion as surface air temperature rises in each scenario, but annual variability and dispersion in precipitation among ensemble members are greater than those in temperature.

Figure 2 shows changes in sea ice. Both coverage and volume decrease associated with temperature rise in each hemisphere. An interesting feature is found in the Northern Hemisphere. Sea ice area and volume decrease abruptly in the last few decades in the 21st century for A1FI and at around 2170 for A1B (Figure 2(b) and 2(d)). This suggests some effects of non-linear response of sea ice or air-sea-ice interaction on sea ice change.

The sea level change due to thermal expansion is calculated by globally integrating the local change of density of sea water relative to a reference state, which is made from average for the PIcntrl experiment and is invariant across all IPCC simulations by the MRI-CGCM2.3. The global average sea level change is a sum of volume change due to thermal expansion and contributions of change of water mass due to water fluxes from land and atmosphere. Since the MRI-CGCM2.3 does not simulate any dynamical change of ice sheets (the model does not have glacier as well), accounting the snow mass over ice sheets into contribution to sea level change is not appropriate. Only the soil moisture and snow mass over the soil land (excluding ice sheets) are accounted for the water flux from land. From the point of view of water budget, water mass over ice sheets is assumed invariant in the model.

Figure 3 shows global average sea level change from thermal expansion in all the IPCC AR4 simulations. The sea level rise ranges from 0.10 m (SRESB1) to 0.15m (SRESA2) at 2100 relative to 2000. This result is slightly higher than previous version (MRI-CGCM2.0, IPCC 2001), but still in the lower limit among other modeling groups. Since the deep ocean temperature increases very slowly, the sea level rise continues even after stabilizing anthropogenic GHG forcing. The additional sea level rise from 2100 to 2300 is 0.10m (SRESB1) to 0.15m (SRESA1B). In the warmer climate, soil moisture increases generally at high latitudes and hence ocean water mass decreases, which leads to a decrease of sea level change about 9%. Also the increase of atmospheric water vapor contributes to descend the sea level several millimeters.
Figure 1  (a) Changes in global and annual mean surface air temperature (degrees centigrade) from the average of 1961-1990 of 20C3M runs. (b) Same as (a) but for precipitation (%). Thin line shows individual runs for ensemble experiments. Black line in (a) shows observed anomaly from the average for 1961-1990 by Jones et al. (2001).

Figure 2  Changes in (a) sea ice coverage ($10^6$ km$^2$) and (b) sea ice volume ($10^{12}$ m$^3$) for the Northern Hemisphere. (c) and (d) are the same as (a) and (b) but for the Southern Hemisphere. Only one member is shown for each ensemble experiment.
4.2 Geographical Response

Figure 4 shows the geographical distribution of the change in annual mean surface air temperature for all 6 SRES scenarios. Temperature rise is greater over continents than over oceans, and greater in the polar regions than in tropical region. These features are consistent with previous studies. In the tropical Pacific, temperature rise is greater in the central-east part than in the western part. Such an anomaly pattern is observed in the El Niño years. The warming patterns among the scenarios are similar although the magnitudes are different.

Figure 5 shows the geographical distribution of the change in annual mean total precipitation. Associated with the El Niño-like response of surface air temperature, precipitation increases in central-eastern tropical Pacific and decreases in western tropical Pacific. Generally, precipitation decreases in the middle latitude arid region and increases middle-high latitude.

The ocean model of the MRI-CGCM2.3 has rigid lid approximation, so the local sea level change is calculated from integration of sea surface pressure gradient. Since the ocean temperature change is not uniform, and furthermore the ocean circulation changes with climate change, the sea level change is non-uniform and reveals a geographical pattern. The geographical distribution of the local sea level change is shown in Figure 6. There is substantially no difference between patterns for all the simulations. The sea level rise is large in the mid-latitude North and South Pacific and the Southern Ocean, and relatively small in the Arctic, North Atlantic and Antarctic Oceans. The largest increases are seen in the east of Japan and Australia, which imply spin-up of the ocean gyre in response to changes in the atmospheric circulation.
Figure 4  Geographical distribution of annual mean surface air temperature change (degrees centigrade) between the periods 1961-1990 and 2071-2100 for (a) A1B, (b) A1T, (c) A1FI, (d) A2, (e) B1 and (f) B2.

Figure 5  Same as Figure 4, but precipitation change (mm day⁻¹).
Figure 6  Geographical distribution of sea level change averaged for 2071-2100 of (a) SRESA1B, (b) SRESB1, (c) SRESA2, and (d) Commit.

5. Summary

We have performed the experiments for all 6 SRES scenarios with some extension to stabilization scenarios proposed for the IPCC AR4. The data sets obtained from the experiments have been submitted to the IPCC Model Output which was established to support the Working Group I component of the AR4 at the PCMDI. Surface air temperature rises about 2.4, 2.1, 3.2, 2.7, 1.7 and 2.0 degrees centigrade for A1B, A1T, A1FI, A2, B1 and B2 scenarios, respectively, between the periods 1961-1990 and 2071-2100. The geographical anomaly response pattern from the global mean in surface temperature is similar to that observed over the Pacific region in the El Niño years. An abrupt change in sea ice is found in a few decades in Northern Hemisphere for some scenarios. This suggests some effects of non-linear response or air-sea-ice interaction on sea ice change. The sea level rise due to thermal expansion is calculated for all the IPCC AR4 simulations. The global average sea level increase ranges from 0.10m (SRESB1) to 0.15m (SRESA2) at 2100 relative to 2000. Geographical response patterns are rather similar among the scenarios although the global means are different.
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Presentations and Publications


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2. Atmospheric and Oceanic Environment Modeling
Ultra-high Resolution Modeling of the Tropical Air-Sea Interaction: Spontaneous Concentration of Cloud Activity in “Planetary” Scale

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Abstract

The space-time structure of cloud convection in a very large-domain, two-dimensional cumulus-resolving model is examined. The model extends 32,768 km in the horizontal direction and is integrated 80 days with surface fluxes from the ocean and a constant body cooling being applied. In the case where the body cooling is stronger in lower troposphere, the activity of cumulus convection spreads in a fairly homogeneous manner. In contrast, in the case where the body cooling is stronger in upper troposphere, the activity of cumulus convection, which is organized in standing oscillation occurring in rather homogeneously at first, becomes gradually concentrated to form a single region. Such spatial structure, which consists of a single, concentrated updraft and a broadly distributed, weak downdraft, corresponds to the spatial structure of “CIFK”, i.e., conditional instability of the first kind. Wave-CISK with strong coupling between cloud and low-level vertical motion is another possibility.

Keywords: WISHE, Cumulus Convection, wave-CISK, CIFK

1. Introduction

Cumulus convection is one of the most important elements in the energy and hydrological cycle of the earth’s atmosphere. But its smallness 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 general circulation model (GCM). Thus every GCM includes cumulus parameterization, which has many uncertainties that arise from our poor understandings on the interaction between large-scale motions and the cumulus clouds. Similar concern also applies to the tropical air-sea interaction (TASI). Usually, the TASI is thought of as a large-scale problem; typical example is El Niño and Southern Oscillation, whose spatial and temporal scales are O (10,000 km) and O (2 years), respectively. Such treatment can partially be justifiable on the grounds that (1) tropical wind system controlling the air-sea exchange processes has large horizontal scale, and that (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 are produced by cumulus convection. The significance of the small-scale components to the TASI problem as a whole is still unclear.

Nakajima (1994) used a two-dimensional cloud model that covers a domain of 16,384 km; its high spatial resolution (2 km) and incorporation of cloud micro-physical processes allows cumulus convection to be represented explicitly, and at the same time the domain is as large as the longitudinal size of the Pacific. If we make an ocean numerical model whose resolution and domain size is similar to the cloud model and conduct coupled experiments, lots of information on the TASI in the wide range of resolution will be collected. Nakajima (1998, 1999, 2000) constructed such fine resolution ocean model and examined its ability to
simulate time evolution of fresh water layer in the tropical ocean caused by convective rainfall.

As an extension of Nakajima (1994), Nakajima (2001) further investigated spontaneous organizations of cloud convection in the cumulus resolving two-dimensional atmospheric model. The results included (1) a large-scale propagating organization of convective activity that emerges in cases with a more realistic radiative cooling, and (2) significant intra-seasonal time scale vacillation of area-averaged precipitation in cases with wind-induced surface heat exchange (WISHE). Nakajima (2002) examined the former type of organization of clouds and interpreted it as a new type of wave-CISK. Nakajima (2003), based on numerical experiments extended up to 160 days, focused on the intra-seasonal time scale vacillation of domain averaged convective activity.

The present paper examines space-time structure of cloud convection in the absence of WISHE feedback. As has been documented in the previous studies mentioned above, propagating organization of convective activity governed by wave-CISK, i.e., the interaction between the heating by clouds and the large-scale atmospheric wave system can emerge in the absence of WISHE. Because wave-CISK-related disturbances have relatively short wavelength (typically about 4000 km) and fast propagation speed (about 20 m/s), their contribution to the temporally averaged rainfall rate was rather weak in previous experiments. However, these results are obtained in relatively short time integration of the models; if larger models are used and integrated for longer duration, wave-CISK-related disturbances may extend to longer wavelength, or other type of large-scale cloud organization may appear. Here, some preliminary results of O (100 days) integration of 32,768 km two-dimensional cloud resolving model will be presented.

2. Design of experiment

The numerical model is a two-dimensional cumulus convection model. Basic equations are the anelastic system, and three-category warm rain parameterized cloud microphysics is included. The computational domain extends 32,768 km horizontally and 22 km vertically. Cyclic boundary condition is assumed in the horizontal direction. The spatial resolution is 2 km in horizontal direction and is about 0.6 km in vertical direction. At the model's lower boundary, heat and moisture fluxes are supplied using the bulk formula assuming that the sea surface with homogeneous temperature exists below. The model troposphere is cooled at a constant rate simulating the effect of radiation very crudely.

In Nakajima (2003), in order to introduce WISHE feedback, a uniform low-level basic "easterly" wind was assumed in the surface flux calculation. In present study, no basic wind is assumed. Furthermore, irrespective of the in situ wind speed in the model's computational grid points, the wind speed in the surface flux calculations is fixed at a constant, homogeneous value, 10 m/s. By that, any WISHE feedback will be switched-off.

It is known (e.g., Nakajima 2002) that, in the absence of WISHE feedback, the space-time structure of convective activity is mainly controlled by the vertical profile of the "radiative" cooling introduced in the model. In this paper, two types of cooling are used; one is "short cooling" where the radiative cooling is stronger in the lower troposphere, and, the other is "tall cooling" where it is stronger in the upper troposphere. The structure of convection may be sensitive to slight difference in experimental setup including initial condition. In order to examine such possible sensitivity, ensemble experiment consisting of 8 members, starting from initial condition slightly different from those for other members, are conducted in some case. The model is integrated for 80 days in each experiment.
3. Results

Figure 1 shows the temporal evolution of rainfall intensity in case with WISHE feedback. As was presented in Nakajima (2003), convective activity is grossly organized in wave number one, eastward propagating structure, although cloud clusters and mesoscale systems are evidently simulated. When WISHE feedback is switched-off, such large-scale organization disappears.

![Temporal evolution of rainfall intensity in the WISHE experiment](image)

Figure 1 Temporal evolution of rainfall intensity in the WISHE experiment. Time goes down.

Figure 2 shows the temporal evolution of rainfall intensity in the case where the WISHE feedback is excluded and the “short” cooling profile is used. Large-scale structure, having O (10,000 km) wavelength, is absent. Still, modulation of cloud activity whose horizontal scale is several hundreds of kilometers develops. The regions of stronger cloud activity is scattered in a fairly homogeneous manner, and each of the active regions of convection moves back and forth slowly, presumably advected by horizontal wind.

Figure 3 shows the temporal evolution of rainfall intensity in the case where the WISHE feedback is excluded and the “tall” cooling profile is used. The structure of convective activity is drastically different between the earlier half and the later half of the experiment; in the first 30 days or so, it is characterized with “mesh-like” pattern originating from fairly homogeneously distributed organization of convective activity that propagates both eastward and westward having the wavelength of about 3,000 km, but, in later time, convective activity occurs almost exclusively in a stationary region with the width of about 2,000 km. Close examination of Figure 3 reveals that the stationary region of concentrated cloud activity develops gradually from wavenumber one stationary modulation of convective activity that appears around t=10 days and steadily develops with time scale of about 20 days.
Figure 2  Temporal evolution of rainfall intensity in the non-WISHE, short-cooling experiment. Time goes down.

Figure 3  Temporal evolution of rainfall intensity in the non-WISHE, tall-cooling experiment. Time goes down.
Figure 4 shows the temporal evolutions of rainfall intensity in the ensemble experiment with the setup common to that used in the case in Figure 3. Although the location or the timing of the development varies due to the slight difference in the initial conditions, gross evolution of the convective activity, i.e., homogeneous, propagating, short wavelength organization in the earlier period and concentration into a stationary narrow region in the later period, are common in all members. This proves that the development of a single, concentrated cloud activity in the large domain is a robust feature in this setup.

Figure 4  Temporal evolutions of rainfall intensity in the non-WISHE, tall-cooling ensemble experiment. The results for all of the 8 runs are shown. Time goes down.
4. Concluding Remarks

The appearance of a single, concentrated cloud activity reminds us of the picture of CIFK originally proposed as a theory to explain the development of tropical cyclones by Kasahara (1961). Today, CIFK is understood to explain the development of individual convective cloud, whose horizontal scale is $O$ (1 km). It is interesting that, if the horizontal scale of the disturbance is taken as that of the characteristic scale of the concentrated cloud activity that appeared in the present experiment, $O$ (10,000 km), the growth time of CIFK disturbance becomes 40 days, which is not very far from the time required for its development. Of course, this coincidence may be fortuitous.

Another interpretation of the appearance of the stationary, concentrated cloud activity is that it is a result of wave-CISK. It can be shown (although not presented here) that parameterized cumulus heatings that have common vertical profile and different magnitude result in quite different behaviors of unstable modes. Namely, stronger coupling between heating and low level vertical motion predicts stationary growing disturbance, while weaker coupling predicts propagating growing disturbance.

In the present experiments with tall cooling, the strength of coupling between vertical motion and cumulus activity may differ for the propagating, short-wavelength structure that dominates in earlier half of the experiment and for the stationary, wavenumber one cloud activity that dominates in the later time. Considering that the cloud activity requires some finite time to respond to the vertical motions, we can expect that the coupling is weaker for non-stationary mode. This is consistent to co-existence of propagating and stationary modes in the present experiment.

If large-scale cloud convection in different setup, where propagating, long wavelength structure is supported, also has some tendency to organize itself in a single stationary concentrated region as exemplified here, then, something of “beat” can occur as the interference between the propagating and stationary structures. This may explain the vacillation that occurred WISHE experiments in Nakajima (2003). The validity of this hypothesis has to be checked by more extensive analysis of the numerical experiments.

Acknowledgement

The author wishes to thank staff members of CGER for providing us with this powerful computing facility. Our software environment was provided from the resources of GFD-DENNOU CLUB.

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3. Geophysical Fluid Dynamics
Passive and Active Scalar Diffusion in Unsteady Stably Stratified Turbulence

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Abstract
Passive and active scalar fluxes in unsteady stratified turbulence are analyzed using the rapid distortion theory (RDT). Solutions of the RDT equations show that viscosity and diffusion excite a 'slowly' oscillating mode in the turbulent diffusion coefficient of the passive scalar. The slow mode becomes dominant even with a small amount of initial potential energy. The results illustrate the importance of unsteady analysis of the unsteady turbulence, which has been often ignored in the previous studies.

Keywords: Stratification, Scalar transport, Passive scalar, Active scalar

1. Introduction

The clarification of the mechanisms of heat and mass transfer in stably stratified turbulence is important in understanding the turbulent transport in the atmosphere and ocean most part of which is composed of the fluid with stable stratification. One of the special characteristics of stratified turbulence is the 'counter-gradient' flux (Komori et al. 1983, Hanazaki & Hunt 1996), which transports heat from the lower cold region to the upper hot region. If this occurs, the vertical turbulent heat flux becomes positive (density flux becomes negative) and the turbulent diffusion coefficient of heat/density often used in the turbulence models becomes negative.

The passive scalar diffusion in this system is important related to the pollution problems, but there have been few fundamental studies which concentrated on the difference between the passive and active (heat or density) scalars. It is true that if the initial conditions and the molecular diffusion coefficients are the same in the passive and active scalars, the turbulent diffusion coefficients also should be the same. Then, our question here is "On which conditions do they become different?"

In the previous studies, Warhaft (1976) showed that turbulent diffusion coefficient for the passive and active scalars becomes different if the cross correlation coefficient between the active and passive scalars is small. However, these discussions in the literature assume stationary turbulence, and the effects of unsteadiness or the initial conditions of turbulence have not been considered.

Numerical simulations have been done by Kaltenbach et al. (1994). They investigated the time development of the cross correlation coefficients between the passive scalar $c$, active scalar $\rho$ and the vertical velocity $u_3$ (i.e., $R_{c3}, R_{\rho3}$ and $R_{c\rho}$) and found that $R_{c\rho}$ reaches a large value ($\geq 0.8$) in a long time even with zero initial correlation.

Nagata and Komori (2001) did experiments for the case of no initial correlation between the passive and active scalar ($R_{c\rho}(t=0)=0$). Time development of $R_{c\rho}$ showed that it reaches a small value, which might be due to the low initial correlation.

In this study we solve the problem of passive scalar diffusion in stably stratified
turbulence as an “initial value problem”, to clarify what determines the subsequent time
development of the fluxes. We use the rapid distortion theory (RDT) based on the linearized
governing equations. It is the only theory that includes the effects of initial conditions. The
results will be useful also in the future numerical and experimental studies in determining the
initial conditions to be used.

2. RDT equations

We consider a fluid with uniform vertical density stratification and with uniform vertical
gradient of a passive-scalar concentration. The governing equations are the Boussinesq
equations for the momentum, equations for the density and passive-scalar advection and the
condition of incompressibility, which can be written as

\[
\frac{\partial \mathbf{u}}{\partial t} + U_j \frac{\partial \mathbf{u}}{\partial x_j} = -\frac{1}{\rho_0} \frac{\partial p}{\partial x_j} - \frac{g}{\rho_0} \rho \delta_{13} + \nu \frac{\partial^2 u_j}{\partial x_j^2}, \tag{1}
\]

\[
\frac{\partial \rho}{\partial t} + U_j \frac{\partial \rho}{\partial x_j} + u_3 \frac{\partial \rho}{\partial x_3} = \kappa \frac{\partial^2 \rho}{\partial x_j^2}, \tag{2}
\]

\[
\frac{\partial c}{\partial t} + U_j \frac{\partial c}{\partial x_j} + u_3 \frac{\partial c}{\partial x_3} = D \frac{\partial^2 c}{\partial x_j^2}, \tag{3}
\]

\[
\frac{\partial \mathbf{u}}{\partial x_j} = 0, \tag{4}
\]

where \( \mathbf{u} = (u_1, u_2, u_3) \) is the perturbation velocity, \( \rho \) and \( c \) are the density and passive scalar
perturbations, \( g \) is the acceleration due to gravity. We substitute the following spectral
decomposition

\[
u_{1}(x,t) = \sum_{k} \hat{u}_{1}(k,t)e^{ik\cdot x}, \quad \rho(x,t) = \sum_{k} \hat{\rho}(k,t)e^{ik\cdot x}, \quad c(x,t) = \sum_{k} \hat{c}(k,t)e^{ik\cdot x}. \tag{5}
\]

into the governing equations (1)-(4) and obtain equations for the spectral components. If we
neglect the nonlinear terms in (1)-(3), we can obtain the RDT (Rapid Distortion Theory)
equations:

\[
\left( \frac{d}{dt} + \nu \mathbf{k}^2 \right) \hat{\mathbf{u}}_3 = \left( \frac{k_3 \cdot k_3}{k^2} - \delta_{13} \right) \hat{\rho},
\]

\[
\left( \frac{d}{dt} + \kappa \mathbf{k}^2 \right) \hat{\rho} = N^2 \hat{u}_3,
\]

and

\[
\left( \frac{d}{dt} + Dk^2 \right) \hat{c} = -\rho \hat{\mathbf{u}}_3,
\]

where \( N \) is the Brunt-Vaisala frequency and \( \gamma \) is the vertical gradient of the mean passive
scalar, both of which are constants independent of \( x_3 \).

3. Turbulent diffusion coefficients

We solve the RDT equations (6)-(8) and obtain the three-dimensional spectra, then
integrate them in the whole spectral space to obtain the fluxes. We assume here that the initial
fluxes of density and passive scalar are zero as in the usual experiments for grid turbulence
and in the previous direct numerical simulations (DNS). We also assume that the turbulence is initially isotropic.

\begin{figure}
\centering
\includegraphics[width=0.5\textwidth]{figure1.png}
\caption{Time development of the viscous/diffusive turbulent diffusion coefficient of density $K_\rho$ for various values of initial energy ratio $PE_0/KE_0$. Solid line: $PE_0/KE_0=0$; short dashed line: $PE_0/KE_0=0.2$; long dashed line: $PE_0/KE_0=0.5$; dash-dotted line: $PE_0/KE_0=1.0$.}
\end{figure}

\begin{figure}
\centering
\includegraphics[width=0.5\textwidth]{figure2.png}
\caption{Time development of the viscous/diffusive turbulent diffusion coefficient of passive scalar $K_\iota$ for various values of initial energy ratio $PE_0/KE_0$. Solid line: $PE_0/KE_0=0$; short dashed line: $PE_0/KE_0=0.2$; long dashed line: $PE_0/KE_0=0.5$; dash-dotted line: $PE_0/KE_0=1.0$.}
\end{figure}

Figures 1 and 2 show the vertical density diffusion coefficient $K_\rho(t)$ and $K_\iota(t)$ with the effect of molecular viscosity and diffusion. Unlike the inviscid case, $K_\rho(t)$ at $PE_0/KE_0=0.5$ ($PE_0$: initial potential energy, $KE_0$: initial kinetic energy) is not zero and persistently negative.
due to the viscous and diffusive effects.

Comparison of Figure 1 with Figure 2 shows that in viscous/diffusive fluid, passive scalar flux ($K_c$) decays much more slowly than the active scalar flux ($K_\rho$), and at large times ($N \geq 5$), $|K_c|$ is much larger than $K_\rho$, when $PE_0/KE_0 > 0$. This suggests that slow modes contained only in $K_c$, and which decay slowly, become dominant at large times.

![Figure 3](image1.png)  
**Figure 3** Time development of the fast mode of $K_c$ for various values of initial energy ratio $PE_0/KE_0$. Solid line: $PE_0/KE_0=0$; short dashed line: $PE_0/KE_0=0.2$; long dashed line: $PE_0/KE_0=0.5$; dash-dotted line: $PE_0/KE_0=1.0$.

![Figure 4](image2.png)  
**Figure 4** Time development of the slow mode of $K_c$ for various values of initial energy ratio $PE_0/KE_0$. Solid line: $PE_0/KE_0=0$; short dashed line: $PE_0/KE_0=0.2$; long dashed line: $PE_0/KE_0=0.5$; dash-dotted line: $PE_0/KE_0=1.0$. 

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Figures 3 and 4 show respectively the contributions from the fast mode (Figure 3) and slow mode (Figure 4) in $K_c$, which was shown in Figure 2.

Comparison of these two figures show that slow mode decays more slowly, and as time proceeds ($Nt>5$) it becomes dominant. This result would be due to the parameters we have used, namely $Pr = \nu / \kappa =6$ and $Sc = \kappa / D = 600$, which satisfy $Sc > 2Pr/(1+Pr)$.

4. Conclusions

We have found that $Kc$ contains a slowly oscillating mode which becomes important even for a small initial density fluctuations $(PE_0/KE_0\sim 0.1)$, and this mode strengthens the counter-gradient passive scalar flux. Since there have been no DNS data which can be compared with the present RDT results, future comparisons with those data would be necessary to quantify the nonlinear effects. We should also consider the effect of mean shear in the mean flow, which usually exists in the real atmosphere and ocean.

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An Aqua-planet Experiment on Structurization of Equatorial Precipitation Activity and Related Software Development toward an Atmospheric General Circulation Model for Terrestrial Planets

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Abstract

An aqua-planet general circulation model (GCM) experiment is performed to examine the effect of the vertical profile of radiative cooling on the structural formation of precipitation activities at around the equator. The purpose is to have further understanding of the possible features of equatorial precipitation activity represented by GCMs with the spatial resolution of climate simulation. By changing the absorption coefficients of infrared radiation, what is really intended is to vary the cumulus heating profile. The results are consistent with the expectation of wave-CISK theory. When the peak of condensation heating is located in the lower troposphere, the grid scale precipitation activity tends to propagate westward, while, when the peak is located in the upper troposphere, it tends to propagate eastward. Composite analyses with reference to precipitation peaks also indicate that the westward structure is conditional instability of the first kind (CIFK) like, while the eastward structure is wave-CISK like. In order to acquire a software environment which enables us to perform GCM experiments like above more easily, we continue to develop a GCM and related softwares covering the parameter ranges of the terrestrial planetary atmospheres. We have packaged "gt4f96io", the network Common Data Form (netCDF) I/O routine, which has been developed in the preceding years, and with the aid of the I/O routine, we started to design a code of three-dimensional primitive equation model which has sufficient readability and portability.

Keywords: GCM, Aqua planet experiment, Wave-CISK, Super cloud cluster, Planetary atmosphere model

1. Background

The purpose of our study is to consider the climatic state of Earth and its variability from the viewpoint of comparative planetary science ranging over Mars, Venus, and Earth. This is also to recognize the circulation structure of Earth from the viewpoint of geophysical fluid dynamics, that is, to reveal possible dynamics embedded under the general circulation of the real atmosphere and to recognize it as one of the realizations in the virtual possible general space of stratified rotating fluid motions. Our activity consists of two major subjects. One is to construct a set of models by which various climatic states of terrestrial planetary atmospheres (the present climatic states of Earth, Venus, and Mars, and the runaway greenhouse state and the ice-covered state) can be simulated, and with which various levels of conceptual frameworks can be connected, for the sake of ensuring the stance of comparative
planetary science and consideration of geophysical fluid dynamics. The other is to reveal possible dynamical structures embedded in the climates of Earth and other planets by performing numerical experiments with some idealistic setups. By reducing the complexity of real situations (e.g. boundary conditions, physical processes, and so on), it may be possible to abstract some fundamental features of the circulation dynamics. In this sense, we refer to our numerical experiments as basic. Those basic numerical experiments are also expected to contribute to providing guidelines for constructing atmospheric models.

As for the basic experiments, we have been performing aqua-planet experiments to investigate the tropical precipitation structures such as Madden-Julian oscillation, super cloud cluster, and cloud cluster. It is a difficult but important problem to search for a desirable representation of small scale vertical convection (cumulus convection for the case of Earth) in general circulation models, especially, in general circulation models with the spatial resolution of climate simulation usage. Cumulus activity in the tropical atmosphere has a strong influence on and almost determines the climate state of Earth. However, the GCMs cannot directly describe the cumulus processes because of the limit of the model spatial resolution. The models are shielded from the issues by introducing some method of cumulus parameterization. However, we still do not understand whether precipitation activities resolved by the model without solving the fine structures of cumulus activity are really reasonable. With those issues in mind, the effect of the vertical profile of radiative cooling on the structural formation of precipitation activities at around the equator is examined.

As for the model and related software development, we are aiming at acquiring a software environment which enables us to perform GCM experiments like described in section 2 and other comparative studies of planetary atmospheres, and to interpret their results more easily. We are proposing a common I/O environment, which can be used by any numerical models. On top of the common I/O environment, we are constructing a hierarchical set of numerical models ranging from shallow water model to full GCM written with a uniform coding rule. In section 3, we describe our present status of the development of the model and related software.

2. Dependency of the Structures of Equatorial Precipitation Activities on the Vertical Profile of Radiative Cooling

2.1 Purpose

It has been argued that the tropical precipitation activities form a hierarchical structure; Madden-Julian oscillation, super cloud cluster, and cloud cluster (e.g. Nakazawa, 1988). By the use of an aqua planet GCM, Hayashi and Sumi (1986) pointed out that super cloud clusters exist as an intrinsic dynamical feature of moist convection of the equatorial atmosphere. Suggested by the results of Hayashi and Sumi (1986), a number of numerical studies concerning tropical precipitation activities have been carried out. From these studies, it is now recognized that precipitation patterns obtained by GCMs strongly depend on implementations of physical processes such as radiation and cumulus parameterization (Lee et al., 2001, Lee et al., 2003). However, the sensitivities of precipitation patterns to the choices of model implementations and the changes of the parameters in them do not seem to be well recognized. For instance, there is a possibility that Hayashi and Sumi (1986) obtained super cloud clusters by chance. The robustness of the results of Hayashi and Sumi (1986) has not been well examined.

It is necessary to perform a thorough parameter study to reveal the diversity of possible precipitation patterns which may be realized under the rather simple setups and with the
simplified physical processes used by Hayashi and Sumi (1986). Such a parameter survey was impossible at the age of Hayashi and Sumi (1986), but now, because of the improvement of computer resources, is within the range of numerical manipulations. One of the important factors causing the diversity of precipitation patterns is the vertical distribution of condensation heating rate. The vertical distribution of heating controls the effectiveness of wave-conditional instability of the second kind (wave-CISK) mechanism (Lau and Peng, 1987) which is, according to Numaguti and Hayashi (1991), essential to the maintenance of super cloud clusters obtained by Hayashi and Sumi (1986). The eastward propagating precipitation activities which is regarded as super cloud clusters is expected to be less coherent and eventually disappear when the vertical distribution of condensation heating rate is not suitable for invoking wave-CISK mechanism.

In the followings, the dependency of precipitation patterns on the vertical distribution of radiative cooling rate is examined with an aqua planet GCM. In order to change the vertical profile of condensation heating we actually vary the radiation property of our model atmosphere, the infrared absorption coefficients. The precipitation activities with various values of infrared absorption coefficients are examined under a simple surface condition; equatorially symmetric and zonally uniform SST.

2.2 Model and Experimental Design

The model used here is GFD Dennou Club AGCM5.3 that is a three-dimensional primitive model which includes simple hydrological and radiation processes. The resolution utilized in our experiment is T42L16. The entire surface is covered with the ocean (aqua-planet configuration) with a fixed value of sea surface temperature (SST). The distribution of SST is equatorially symmetric, and zonally uniform. The radiation scheme is a band model with four bands; one is for dry air and three are for water vapor. In this study, the absorption coefficient for the dry air band is changed, which alters the vertical distribution of convective heating rate. As for the cumulus parameterization, moist convective adjustment scheme or Kuo scheme is used. Kuo scheme is adopted also by Hayashi and Sumi (1986). It is recognized that, because of its formulation, Kuo scheme tends to produce wave-CISK like structures.

2.3 Results for Kuo Scheme

Figure 1 shows the zonally averaged vertical distributions of radiative cooling rate and condensation heating rate obtained by experiment using Kuo scheme. As the absorption coefficient for dry air band increases, radiative cooling rate in the lower troposphere increases, and, as a result, the level with maximum condensation heating rate moves downward. Contrary, as absorption coefficient for dry air band decreases, radiative cooling rate in the upper troposphere increases, and, as a result, the level of maximum condensation heating rate moves upward. Hereafter, the experiment with the largest condensation heating rate in the lower troposphere is referred to as run K-LC and the experiment with the largest condensation heating rate in the upper troposphere as run K-UC.
Figure 1  Zonally averaged vertical distributions of radiative cooling rate (left panel) and condensation heating rate (right panel) at the equator. Red line, black line, yellowish-green line, and light blue line show the results of the runs in which absorption coefficient for dry air band are decreased in this order. Unit is [K/s].

In run K-LC, westward propagating precipitation areas with a scale of grid size appear clearly in longitude-time section (Figure 2, left panel) and zonal wavenumber-frequency power spectra (Figure 3, left panel). The phase speed of the precipitation areas is about 7 m/sec, which is almost equal to zonally averaged background zonal wind. Figure 4 (left panel) shows the composite circulation structure with reference to precipitation peaks in the longitude-height section. The composite structure of convection is upright; no phase slant of temperature and zonal wind can be observed. In this sense, the structure of grid scale precipitation areas is not wave-CISK but a simple direct moist convection, or in other words, conditional instability of the first kind (CIFK). The westward motion of precipitation areas is regarded as advection of CIFK structure by the background wind.

In run K-UC, eastward propagating precipitation areas with a scale of grid size can be clearly observed in longitude-time section (Figure 2, right panel) and in zonal wavenumber-frequency power spectra (Figure 3, right panel). Their phase speed is about 23 m/s. Composite structure (Figure 4, right panel) shows the westward phase slant of temperature and wind in longitude-height section. This result is consistent with the structure expected by wave-CISK theory.

Figure 2  Temporal variation of equatorial precipitation. (Left) run K-LC, (right) run K-UC. Unit is [W/m²].
Figure 3  Zonal wavenumber-frequency power spectra of equatorial precipitation. (Left) run K-LC, (right) run K-UC.

Figure 4  Composite structure with reference to precipitation peaks at the equator. Anomalies from zonal mean values are shown. Left panel and right panel show the results of run K-LC and run K-UC, respectively.

2.4 Results for Adjustment Scheme

Figure 5 shows zonally averaged vertical distributions of radiative cooling rate and condensation heating rate obtained by experiment using convective adjustment scheme. Similar to the results of the experiment with Kuo scheme, with increased absorption coefficient for dry air band, the level of maximum radiative cooling rate moves downward. Hereafter, the experiment with largest condensation heating rate in lower troposphere is referred to as run A-LC and the experiment with largest condensation heating rate in upper troposphere as run A-UC.
Figure 5  Same as Figure 1 but for adjustment scheme.

The results with adjustment scheme are shown in Figures 6 and 7. In both of the runs A-LC and A-UC, precipitation areas with a scale of grid size propagate westward at the equator. In run A-UC, envelopes of grid scale precipitation activities seem, although ambiguous, to exist and can be recognized to propagate eastward.

Figure 8 (left panel) shows the composite structure with reference to the precipitation peaks of westward propagating precipitation activities in run A-UC. Phase slant of temperature and wind fields are not observed. On the other hand, in the composite structure with reference to the precipitation peaks of eastward propagating precipitation activities (Figure 8, right panel), there exist westward phase slant of temperature and wind. However, slant of phase is not so clear as that observed in the result of run K-UC.

Figure 6  Temporal variation of equatorial precipitation. (Left) A-LC, (right) A-UC. Unit is [W/m²].
Figure 7  Zonal wavenumber-frequency power spectra of equatorial precipitation. (Left) A-LC, (right) A-UC.

Figure 8  Composite structures of precipitation structures obtained in run A-UC. Left panel and right panel show composite structures with reference to precipitation peaks of westward propagating structures and eastward propagating structures, respectively. Anomalies from zonal mean values are shown.

2.5 Summary

In our model, GFD Dennou Club AGM5.3, coherent eastward propagation of grid scale precipitation activities are rarely observed compared to those obtained by Hayashi and Sumi (1986) and Numaguti and Hayashi (1991). The tendency of the appearance of coherent eastward motion of precipitation areas does agree, especially for the cases with Kuo parameterization, with the expectation of wave-CISK theory. Top heavy condensation heating profile is favorable for the appearance of wave-CISK coupling. However, our GCM tends to produce, in general, westward propagating structures which are considered to be areas of simple moist convection, or in other words, maintained by CIFK, advected by the background wind. In most of the runs, coupling of waves with moist convection does not seem to be evident. It is curious that westward propagating precipitation activities with CIFK-like structure dominate even in runs with Kuo scheme, since Kuo scheme is considered to be favorable for wave-CISK mechanism and is used in the above mentioned previous studies where coherent eastward motion of grid scale precipitation areas is quite evident. As for the
reasons why wave-CISK mechanism does not work effectively in our model compared to the previous works, we speculate that the vertical profile of radiative cooling rate and/or SST distribution are different from those of previous works. However, at the present stage, parameter study with regards to the radiation profile, and even SST profile, is not enough to figure out the tendency of the appearance of precipitation patterns.

Our results show that vertical distribution of condensation heating rate strongly influences on the appearance of tropical precipitation patterns. The condensation heating rate profiles utilized in our experiment are in the range of those obtained in several GCMs (Lin et al., 2004). Our results suggest that the appearance of tropical precipitation patterns can change drastically with the difference of model implementation of the cumulus process.

3. Development of Models and Related Softwares

We have been trying to develop a set of models and related softwares by which we can perform GCM experiments like described in section 2 and other comparative studies of planetary atmospheres more easily. GFD-Dennou-Club planetary atmosphere model (DCPAM) is a symbolic name of a set of hierarchical models which consist of vertical one dimensional model, north-south one dimensional model, two dimensional meridional model, two dimensional longitudinal model, two dimensional spherical horizontal model, and three dimensional global model. DCPAM is aimed to cover the parameter range of planetary atmospheres including Earth, Mars, and also the surface atmospheric layer of Jupiter, and to serve for consideration of the circulation structures of those objects from the viewpoint of geophysical fluid dynamics. We started to design and preliminary implementation test for DCPAM from last year. With above mentioned aims in mind, we emphasize the following four points on the coding structure of the models:

- Flexibility of changing the schemes implemented in the models
- Portability of physical processes as software modules which can be commonly used among the models
- Readability of the source code
- Portability of data which can be easily transferred among the computing platforms

Figure 9 shows software structure related to DCPAM. Data format used by DCPAM is based on network Common Data Form (netCDF), which enables data transparent through network. As for the data structure, we adopt gtool4 netCDF convention (Toyoda et al., 2004) which is an extension of Cooperative Ocean/Atmosphere Research Data Service (COARDS) netCDF convention, that embodies a self descriptive data structure for analysis and visualization. We have partly rebuilt and reorganized the Fortran 90 library for gtool4 netCDF data I/O into “gt4f90io” package (Morikawa et al., 2004). With the aid of the I/O routines provided by gt4f90io, we started to design a coding style for the set of hierarchical models. DCPAM is to be developed on the frame of Hierarchical Spectral Models for GFD (SPMODEL; Takehiro et al., 2004) which utilizes basic spectral transfer functions provided by ISPACK (Ishioka, 2002). In SPMODEL, suppression of dummy indices is realized by supplying Fortran 90 modules for spectrum transformation and spatial differentiation, which enable us to greatly improve readability of the source code. For analyzing and visualizing data modules of Dennou Ruby Project (e.g. Gphys, RubynetCDF, Dennou Ruby Project, 2004) developed with object oriented script language Ruby are to be utilized. We are now developing a three-dimensional primitive atmosphere model with the basis of a spherical shallow water spectrum model which was developed last year.
Figure 9  Software structure realted to DCPAM.

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Direct Numerical Simulation of Turbulent Transfer in Convective Boundary Layer beneath the Gas-Liquid Interface

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Abstract
To parameterize the CO₂ flux across air-sea interface exhaustively, we examined the effect and nature of the heat convection developing beneath the air-sea interface by applying the three-dimensional Direct Numerical Simulation (DNS). A laboratory experiment of Karusudani et al. (1995) presented that heat and mass transfer coefficient, Nusselt number, in heat convection beneath gas-liquid interface is twice as large as that in convection over a solid wall. This effect in the East China Sea corresponds to that of the wind of 3-4ms⁻¹ over the sea. It means that the heat convection developing beneath the sea surface significantly contributes to the uptake of CO₂ into the ocean and neglection of this effect lead to a significant underestimation of the ocean flux. The result of present study shows almost the same result obtained from the laboratory experiment of Karusudani et al.

Keywords: Direct Numerical Simulation (DNS), Heat convection, Mass transfer, Liquid turbulence

1. Introduction
Turbulent structure and transport process beneath the gas-liquid interface is important in air-sea interaction and in many industrial operations. This is important outstandingly in the uptake of atmospheric greenhouse gases into ocean, since the uptake rate of the slightly soluble gases like CO₂ between the atmosphere and ocean is controlled by the liquid-side mass transfer resistance in a very thin film of the order of 10-100 μm beneath the interface and so by the liquid-side turbulence structure.

The liquid-side turbulence near the surface controlling mass transfer between air and sea is induced by wind shear and buoyancy. Mass transfer across the gas-liquid interface in the first case has been investigated extensively by the theoretical approaches, the field measurements (McGillis et al., 2001), laboratory experiments (Komori et al., 1993; Wanninkhof and Bliven, 1991) and numerical simulations (Pan and Banerjee, 1995). However, the effects of buoyancy-induced turbulence have hardly been discussed at all. So the behaviors of buoyancy-induced turbulence just beneath the liquid surface have not been well-known quantitatively.

The cold wind blowing over the warm ocean causes the latent heat and sensible heat fluxes at the sea surface. Whereby the temperature of the sea surface is cooler than that of undersea, and then the turbulent heat convection arises beneath the sea surface, which brings CO₂ in the ocean. In the oceanic surface convection layer, convective eddies impinging on the liquid surface cause a slight deflection of the surface and the latter, in turn, produces an
excess pressure and pushes back the eddies. It results in the wave motion at the liquid surface, dampening the turbulent vertical motion and redistributing its energy to the horizontal one near the surface. This might cause a substantial difference of turbulence structure with that in the convection over a solid surface. A laboratory experiment of Karasudani et al. (1995) presented that heat and mass transfer coefficient, Nusselt number, in heat convection beneath the gas-liquid interface is twice as large as that in convection over a solid wall. This contribution in the East China Sea corresponds to the effect of the wind of 3–4 ms\(^{-1}\) over the sea.

The purpose of this study is to simulate the turbulent convective boundary layer beneath the gas-liquid interface by the three dimensional Direct Numerical Simulation (DNS), and compare the present results with that of Karasudani et al.

2. Direct Numerical Simulation

2.1 Flow Configuration

Numerical object considered is the three-dimensional turbulent heat convection beneath a cooling flat free-slip boundary, since we can consider free surface with no shear stress as free-slip wall. The governing equations are the incompressible Newtonian Navier-Stokes equation with the Boussinesq approximation and mass and heat conservation equations,

\[
\nabla \cdot \mathbf{u} = 0, \tag{1}
\]

\[
\frac{\partial \mathbf{u}}{\partial t} + (\mathbf{u} \cdot \nabla) \mathbf{u} = -\frac{1}{\rho_0} \nabla p + \beta(T - \overline{T})g\delta_z + \nu \nabla^2 \mathbf{u}, \tag{2}
\]

\[
\frac{\partial T}{\partial t} + (\mathbf{u} \cdot \nabla) T = \kappa \nabla^2 T, \tag{3}
\]

where \(\rho_0\), \(g\), \(\nu\), \(\kappa\) and \(\beta\) are the reference density, the gravity acceleration, kinetic viscosity, thermal diffusivity and coefficient of thermal expansion, respectively. \(\overline{T}\) is the horizontal averaged temperature. The buoyancy term operates only in the vertical direction.

The computational domain is shown schematically in Figure 1 and the initial condition is stably stratified as shown in Figure 2. The temperature of the upper boundary, \(-\Delta T\), is cooler than the initial temperature of fluid beneath the free surface so that the upper region in the computational domain is to be unstable. Random perturbation is added to the temperature and the velocity fields to develop turbulent flow.

The boundary condition is set up to free-slip condition for the upper boundary, non-slip condition for the bottom one and periodic condition in horizontal directions. The aspect ratio of computational domain is chosen to be 6 that is large enough that periodic boundary condition doesn’t disturb the large-scale flow. The Prandtl number is 1.0.
2.2 Computational Method

A finite-difference method is adopted to approximate the governing equation. All spatial derivatives in the governing equations are discretized by the second order central-difference on a staggered grid. The advective terms are advanced by the explicit second-order Adams-Bashforth method and viscous terms by the implicit second-order Crank-Nicholson method. The equations are developed following the fractional step method, and a Poisson equation is solved directly, using the fast Fourier transforms in the two horizontal directions and tri-diagonal matrix inversions in the normal direction. The numbers of grid points are 192 in the horizontal directions and 128 in the normal direction. The calculation is repeated until the flow is statistically quasi-steady state.

3. Numerical Results and Discussions

Figure 3 depicts the vertical distribution of time-space averaged temperature $\bar{T}$. The height is normalized as the depth of convective layer $z$, which is defined as the height of the heat flux equal to 0. The averaged temperature is linear distribution and unstable state near the
free surface. The Rayleigh number is 13771. This value is so low that the flow is laminar state in the case of the two wall-bounded convection like Rayleigh-Benard convection. However, as shown in Figure 4, which denotes the horizontal distributions of simultaneous temperature at \( z = 0.1z_i \) and \( z_i \), these distributions are not orderly laminar states.

![Figure 3](image1.png)

**Figure 3** Vertical distribution of the averaged temperature.

![Figure 4](image2.png)

**Figure 4** Instantaneous temperature distribution at \( z = 0.1z_i \) (left) and \( z = z_i \) (right).

In Figure 5 and 6 vertical profiles of the vertical and horizontal velocity variances, \( u^2 \) and \( w^2 \), which are normalized as the convective velocity scale \( w_* = g \beta F_{wall} z_i \), are compared with the results of laboratory experiment of Karasudani et al., and the field measurements of convection over the solid surface which is the results of atmospheric convective layer observed by Caughey and Palmer (1979). The height, where the vertical variance is maximum, of the present DNS is respectively located higher than those of experiment and measurements, because the Rayleigh number of the present DNS is much less than those of experiment and measurements. But the correlation among them is high. The
horizontal variance is, on the other hand, not in good agreement with that of laboratory experiment. Compared with that of laboratory experiment, that of present result is large near the outer edge of convective boundary layer.

Figure 5  Vertical distribution of normalized vertical variance. The solid curve is from the present DNS, + is from the laboratory experiment of Karasudani et al. (1995).

Figure 6  Vertical distribution of normalized horizontal variance

Nusselt number, Nu, is generally described as a function of Rayleigh number, Ra, where Nu is defined by $h z_i / \lambda$ and Ra is $g \beta \Delta T z_i^3 / \nu \kappa$, and $h$, $\lambda$, and $\Delta T$ denote the heat
transfer coefficient, the heat conductivity and the temperature difference between the free surface and the outer edge of the boundary layer, respectively. It is represented by

\[ \text{Nu} = c(Ra)^{1/3}. \] (4)

The value of coefficient \( c \) of the present DNS is 0.33 and those of the convection over the solid surface and Karasudani et al. are 0.15 and 0.29, respectively. The coefficient 0.33 is close to that of Karasudani et al., and twice as large as that for solid surface.

4. Conclusion

We applied 3-D Direct Numerical Simulation to the heat convection beneath the free surface, and showed the turbulent statistics. The vertical velocity variance is qualitatively agreement with those of laboratory experiment and field measurements, but on the other hand the horizontal velocity variance is not in agreement with that of laboratory experiment, especially outside of the convective layer.

The dimensionless heat transfer coefficient i.e., Nusselt number of the present DNS is twice as large as that of the convection over the solid wall, which is in good agreement with that of laboratory experiment.

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Merger of Quasi-Geostrophic Ellipsoidal Vortices and Refinements on the Ellipsoidal Moment Model

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Abstract
Refrinement on the quasi-geostrophic vortex model are proposed. Each coherent vortex is modeled by a slender spheroid. The dynamics of N spheroidal vortices is a canonical Hamiltonian system of 2N degrees of freedom. We adapt the Gaussian integration method in the evaluation of the mutual energy integrals, instead of the moment approximation used in the previous model. The accuracy of integration is improved drastically and the refined model can circumvent the singular behavior of a counter-rotating vortex pair.

Keywords: Quasi-Geostrophic vortex model, Hamiltonian system, Gaussian integration method, Refined model

1. Introduction

In geophysical flows, isolated three-dimensional vortices keep their identity for long time. Their interactions, especially mergers of co-rotating vortices, dominate the dynamics of the geophysical turbulence. The continuous dynamical system can be reduced to a system of discrete vortices by constructing a vortex-based model of turbulence.

Miyazaki et al. (2000) developed a simple vortex model, in which each coherent vortex is modeled by a slender prolate spheroid, i.e., a wire-vortex, which is specified by its location (X, Y, Z), its length 2l and its orientation (θ, φ). The dynamical equations were derived systematically, following the procedure of Hamiltonian moment reduction. The system of N interacting wire-vortices has 2N degrees of freedom. Later, Miyazaki et al. (2001) developed an ellipsoidal moment model, by extending the wire moment model. Each vortex is modeled by an ellipsoid of uniform potential vorticity embedded in a 'locally uniform shear field' induced by other vortices. The ellipsoidal vortex is specified by its location (X, Y, Z), its shape (semi axis-length) (α, β, γ) and its orientation (Euler angles) (θ, φ, ψ). The degree of freedom of N interacting vortices is 3N.

Miyazaki et al. (2002) checked the validity of the ellipsoidal moment model, by performing direct numerical simulations based on the CASL-algorithm. D.G. Dritschel et al. (1997) investigated the interaction of two co-rotating vortices placed on two horizontal planes of different vertical height. It was found that the critical distance of merger of co-rotating vortices is estimated by the ellipsoidal moment model fairly well.

Miyazaki et al. (2003) investigated the interaction of two counter-rotating vortices of the same strength and shape (vortex dipole) based on the ellipsoidal moment model. The moment model predicts singular behavior, i.e., one (or both) of tall vortices with α/γ < 2/3, is (are) stretched infinitely when they are placed within a certain critical distance initially. The computation based on the model stops soon after such singular behavior is detected. In their CASL simulations, two counter-rotating vortices emit filaments vigorously, in the parameter region where the moment model predicts singular behavior. The dipole structure remains robustly, even if the two counter-rotating vortices are in touch initially. These results show
that the ellipsoidal moment model should be refined.

The aim of this paper is to develop a refined model. First, we introduce the Gaussian integration method in evaluating the integrals representing the mutual interaction energy between two wire-vortices. The accuracy of integration is highly improved, and the refined model becomes very robust, circumventing the singular behavior of the previous model based on the moment approximation. Next, we introduce a realistic relation between the inclination angle and the vortex-slenderness parameter, assuming that each vortex is a prolate spheroid of finite aspect ratio. This refinement reinforces the ability of the wire model in representing a fatter vortex quantitatively.

In Section 2, we show the derivation of the refined model. We consider the interaction of a counter-rotating vortex pair in Section 3, where we show that the refined model is quite robust. In Section 4, we investigate mutual interactions between two co-rotating vortices. The critical distance of merger (and the occurrence of chaotic motion) is investigated in detail. The predictions of the refined model are in good agreement with the results of the CASL-computations. The last section is devoted to a brief summary.

2. Refinements on the Vortex Model

We propose two refinements of the vortex model, after reviewing the previous moment model briefly. The first and major refinement is about the approximation of the mutual interaction energy, and the second is the improvement of the approximation of the self-energy.

2.1 Moment Model

We assume that the potential vorticity is distributed along N wire-vortices. The strength of the i-th vortex is denoted by $\Gamma_i$, which is the integration of the potential vorticity. The strength of each vortex $\Gamma_i$ is time-independent, because the potential vorticity is conserved.

Figure 1 illustrates the variables that specify wire-vortices. The length of the i-th wire is $2l_i$, and it is tilted from the vertical axis (taken as the z-axis) by $\theta_i$. Its orientation in the x-y plane is specified by the angle $\phi_i$. The center of the i-th vortex is located at $(X_i, Y_i, Z_i)$. Because the fluid motion is confined in each horizontal plane (under the quasi-geostrophic approximation), there are two Casimir invariants $Z_i$, $zh_i = l_i \cos \theta_i$. The latter denotes the half-height of the i-th vortex. The number of variables is reduced to 4N and the degree of freedom is reduced to 2N.

![Figure 1 Variables specifying wire-vortices.](image-url)
The wire-vortex model is most elegantly formulated using the canonical variables \(X_i, Y_i, S_i\) and \(\phi_i\). Here \(S_i\) is related to \(l_i\) and \(\theta_i\) by

\[
S_i = \frac{1}{10} \hat{\Gamma}_i z h_i^2 \tan^2 \theta_i.
\]

(1)

The equation of motion are written as,

\[
\hat{\Gamma}_i \frac{dX_i}{dt} = \frac{\partial H}{\partial Y_i},
\]

(2)

\[
\hat{\Gamma}_i \frac{dY_i}{dt} = -\frac{\partial H}{\partial X_i},
\]

(3)

\[
\frac{dS_i}{dt} = \frac{\partial H}{\partial \phi_i},
\]

(4)

\[
\frac{d\phi_i}{dt} = -\frac{\partial H}{\partial S_i}.
\]

(5)

If we introduce new variables \(x_i = l_i \sin \theta_i \cos \phi_i\) and \(y_i = l_i \sin \theta_i \sin \phi_i\) that represent the \(x\)- and \(y\)-coordinates of the top point \((X_i + x_i, Y_i + y_i, Z_i + z h_i)\) relative to the center of the \(i\)-th vortex, the latter two equations are re-written in a more familiar form:

\[
\frac{\hat{\Gamma}_i}{5} \frac{dx_i}{dt} = \frac{\partial H}{\partial y_i},
\]

(6)

\[
\frac{\hat{\Gamma}_i}{5} \frac{dy_i}{dt} = -\frac{\partial H}{\partial x_i},
\]

(7)

We integrate these equations (2), (3), (6) and (7) numerically in the following sections, after the actual form of the Hamiltonian is given.

The Hamiltonian consists of the self-energy and the mutual interaction energy:

\[
H = \sum_{i=1}^{N} H_{si} + \sum_{(i,j)}^{N} H_{mj}.
\]

(8)

The second summation of the mutual energy is taken for all pairs of vortices (\(N (N-1)/2\) pairs). We must specify these two parts of the Hamiltonian, appropriately. Our previous choice for the self-energy was,

\[
H_{si} = \frac{3}{20 \pi h_i} \log^2 \left( \frac{2 - \log \theta_i}{\log \theta_i} \right).
\]

(9)

with the slenderness parameter

\[
\varepsilon_i = \varepsilon_{\theta_i} \cos^{-1/2} \theta_i.
\]

(10)

This expression is consistent with the self-rotation rate:

\[
\Omega_{si} = -\frac{\partial H_{si}}{\partial S_i} = \frac{3}{8 \pi h_i} \left( \cos \theta_i \right)^{3/2} \left[ 2 (\log 2 + \log \varepsilon_{\theta_i}) - 3 \log (\cos \theta_i) - 3 \right].
\]

(11)
These are correct only at the lowest order of \( \varepsilon \alpha \), i.e., the terms higher than \( O(\varepsilon \alpha) \) are neglected. The slenderness parameter \( \varepsilon \) is linked with the aspect ratio of a prolate spheroid \( \alpha/\gamma \). We will describe an attempt to improve the order of the approximation below.

The exact mutual interaction energy is given by the double integral:

\[
H_{\mu\nu} = \frac{9}{64\pi l_0^2 I_j^3} \int_i \int_j \frac{(l_i^2 - s_i^2)(l_j^2 - s_j^2)}{|R_i - R_j|} \, ds_i \, ds_j,
\]

\[
R_{i(j)} = (X_{i(j)} + s_{i(j)} \sin \theta_{i(j)} \cos \phi_{i(j)}, Y_{i(j)} + s_{i(j)} \sin \theta_{i(j)} \sin \phi_{i(j)}, Z_{i(j)} + s_{i(j)} \cos \theta_{i(j)}).
\]

The vorticity distribution along the i-th wire-vortex should be proportional to \( l_i^2 - s_i^2 \), so that it remains straight. The appearance of this factor \( l_i^2 - s_i^2 \) is also consistent with the fact that a tilted spheroidal vortex rotates rigidly. In order to close the problem, Miyazaki et al. (2000), introduced an approximate formula, in which the above integral was expressed by the moments up to second order. They expanded the inverse of the distance between two vortices \( |R_i - R_j|^{-1} \), assuming that the distance was much larger than the length of the wire vortices:

\[
\frac{1}{|R_i - R_j|} = \frac{1}{R_y}
\]

\[
- \frac{1}{R_y} [(X_i - X_j)(s_i \sin \theta_i \cos \phi_i - s_j \sin \theta_j \cos \phi_j)
+ (Y_i - Y_j)(s_i \sin \theta_i \sin \phi_i - s_j \sin \theta_j \sin \phi_j)
+ (Z_i - Z_j)(s_i \cos \theta_i - s_j \cos \theta_j)]
\]

\[
- \frac{1}{2R_y} [(s_i \sin \theta_i \cos \phi_i - s_j \sin \theta_j \cos \phi_j)^2
+ (s_i \sin \theta_i \sin \phi_i - s_j \sin \theta_j \sin \phi_j)^2
+ (s_i \cos \theta_i - s_j \cos \theta_j)^2]
+ \frac{3}{2R_y} [(X_i - X_j)(s_i \sin \theta_i \cos \phi_i - s_j \sin \theta_j \cos \phi_j)
+ (Y_i - Y_j)(s_i \sin \theta_i \sin \phi_i - s_j \sin \theta_j \sin \phi_j)
+ (Z_i - Z_j)(s_i \cos \theta_i - s_j \cos \theta_j)]^2,
\]

where \( R_y \) is the distance between the center of the i-th vortex and that of the j-th vortex

\[
R_y = \sqrt{(X_i - X_j)^2 + (Y_i - Y_j)^2 + (Z_i - Z_j)^2}.
\]

The integrations with respect to \( s_i \) and \( s_j \) yielded,
\[
H_{mj} = \frac{\hat{T}_{i} \hat{T}_{j}}{4\pi R_{ij}} \left\{ 1 - \frac{I_{i}^{2} + I_{j}^{2}}{10 R_{ij}^{2}} \right\} \\
+ \frac{3}{10 R_{ij}^{4}} \left[ (X_{i} - X_{j})^{2}(x_{i}^{2} + x_{j}^{2}) + (Y_{i} - Y_{j})^{2}(y_{i}^{2} + y_{j}^{2}) + (Z_{i} - Z_{j})^{2}(zh_{i}^{2} + zh_{j}^{2}) \right] \\
+ 2(X_{i} - X_{j})(Y_{i} - Y_{j})(x_{i}y_{i} + x_{j}y_{j}) \\
+ 2(Y_{i} - Y_{j})(Z_{i} - Z_{j})(y_{i}zh_{i} + y_{j}zh_{j}) \\
+ 2(Z_{i} - Z_{j})(X_{i} - X_{j})(zh_{i}x_{i} + zh_{j}x_{j}). 
\] (16)

Miyazaki et al. (2003) noticed that both the ellipsoidal- and wire-moment models predict singular behavior, when two tall counter-rotating vortices are located within a certain critical distance. One (or both) of vortices is (are) stretched infinitely. This is a fatal drawback of the model, because the model-computation stops whenever it occurs. Their direct numerical simulations (CASL), however, showed that these singularities are false predictions of the moment models. This is mainly due to the inaccurate approximation of the mutual energy integral by the moment model, when two vortices are placed close each other.

2.2 Gaussian Integration Method

The basic idea of improving the accuracy of the mutual energy integration is to utilize the Gaussian quadrature formulae. We notice that the essence of the mutual energy integral is approximated by a discrete summation:

\[
\int_{-1}^{1} (1 - s^{2}) f(s) ds = \sum_{i=1}^{2M+1(2M)} w_{i} f(s_{i}). 
\] (17)

Here, the weight \( w_{i} \) and the abscissas \( s_{i} \) are chosen so that the above formula is exact for the polynomials up to \( 4M(4M-2) \) order. If we take \( s_{i} \) symmetrically, the odd order polynomials give no contribution. We need \( 2M+1 \) points when the center (origin) is included and \( 2M \) points when the center (origin) is excluded. We must solve the coupled nonlinear equations (for the odd \((2M+1)\)-case)

\[
\sum_{i=1}^{2M+1} w_{i} s_{i} = \int_{-1}^{1} (1 - s^{2}) s^{2m} ds \]

\[= \frac{2}{2m+1} - \frac{2}{2m+3} \]

\[= \frac{4}{(2m+1)(2m+3)} \] (18)

for \( m=0, 1, 2, \ldots, 2M \), in order to determine the values of \( w_{i} \) and \( s_{i} \). This task is easily
accomplished for small $M$. Further, if we introduce $\tilde{w} = w \times \frac{3}{4}$, the mutual energy integral can be approximated as,

$$H_{mj} = \frac{\Gamma_1 \Gamma_2}{4\pi} \sum_{m=1}^{2M+1} \sum_{n=1}^{2M+1} \frac{w_m w_n}{|R_{mj} - R_{jn}|}.$$  \hspace{1cm} (19)

for example, two points approximation can be written down as,

$$\tilde{s}_{1,2} = \begin{cases} \frac{1}{\sqrt{5}} \\ \frac{1}{\sqrt{5}} \end{cases}.$$  \hspace{1cm} (20)

$$\tilde{w}_{1,2} = \frac{1}{2}.$$  \hspace{1cm} (21)

and

$$H_{mj} = \frac{\Gamma_1 \Gamma_j}{4\pi} \sum_{m=1}^{2} \sum_{n=1}^{2} \frac{w_m w_n}{|R_{mj} - R_{jn}|} = \frac{\Gamma_i \Gamma_j}{16\pi} \sum_{m=1}^{2} \sum_{n=1}^{2} \frac{1}{|R_{mj} - R_{jn}|}.$$  \hspace{1cm} (22)

with

$$R_{i1} = \begin{pmatrix} X_i - \frac{x_i}{\sqrt{5}} \\ Y_i - \frac{y_i}{\sqrt{5}} \\ Z_i - \frac{zh_i}{\sqrt{5}} \end{pmatrix}.$$  \hspace{1cm} (23)

$$R_{i2} = \begin{pmatrix} X_i + \frac{x_i}{\sqrt{5}} \\ Y_i + \frac{y_i}{\sqrt{5}} \\ Z_i + \frac{zh_i}{\sqrt{5}} \end{pmatrix}.$$  \hspace{1cm} (24)
\[ R_{j1} = \begin{pmatrix} X_j - \frac{x_j}{\sqrt{5}} \\ Y_j - \frac{y_j}{\sqrt{5}} \\ Z_j - \frac{zh_j}{\sqrt{5}} \end{pmatrix}, \quad (25) \]
\[ R_{j2} = \begin{pmatrix} X_j + \frac{x_j}{\sqrt{5}} \\ Y_j + \frac{y_j}{\sqrt{5}} \\ Z_j + \frac{zh_j}{\sqrt{5}} \end{pmatrix}, \quad (26) \]

Similarly the discrete approximation of three-points can be written down with the following \( \hat{s}_{m} \) and \( \hat{w}_{m} \):

\[ \hat{s}_{1,2,3} = \begin{pmatrix} -\frac{3}{\sqrt{7}} \\ 0 \\ \frac{3}{\sqrt{7}} \end{pmatrix} \quad \hat{w}_{1,2,3} = \begin{pmatrix} \frac{7}{30} \\ \frac{8}{15} \\ \frac{7}{30} \end{pmatrix}. \quad (27) \]

We have checked the accuracy of the approximation up to the seven points discretization to find out the three points discretization provides satisfactory accuracy.

2.3 Refinement on the Self-energy: Spheroidal Vortex Approximation

In this subsection, we attempt to improve the approximation of the self-energy. The previous formulae (9) discards the terms higher than \( O(\epsilon^0) \), which prevents us from investigating fatter vortices quantitatively. A natural idea is to identify each vortex by a spheroid with the aspect ratio \( \hat{\delta} = \alpha / \gamma \). Indeed, Miyazaki et al. (2000) suggested such possibility. They were, however, unable to construct consistent formulae of the self-energy and the self-rotation rate. An intuitively promising choice of the self-energy (the subscript \( i \) is omitted hereafter)

\[ H_s = \frac{3 \Gamma}{20\pi \sqrt{1 - \epsilon}} \left[ \log(1 + \sqrt{1 - \epsilon}) - \log \epsilon \right], \quad (28) \]

with the self-rotation rate of a tilted spheroid of the aspect ratio \( \epsilon = \tanh \hat{\xi} \).
\[ \Omega_s = \frac{3\Gamma}{16\pi\varepsilon l^3} \cosh\xi \sinh^2\xi \left\{ \log\left(\frac{\cosh\xi + 1}{\cosh\xi - 1}\right) (3\cosh^2\xi - 1) - 6 \cosh\xi \right\}. \] (29)

failed. This was mainly because the relation between \( \hat{\varepsilon}, \theta \) and \( \hat{\varepsilon_0} \) (constant)

\[ \hat{\varepsilon} = \hat{\varepsilon_0} \cos^{3/2} \theta, \] (30)

was insufficient. Here, the slenderness parameter \( \hat{\varepsilon_0} \) was thought to be the aspect ratio of the spheroidal vortex when it stood up vertically without change of its volume \( 4\pi\alpha^2\gamma/3 \) and its half-height \( zh \) then the above relation (30) was derived under two geometrical assumptions that the semi-major axis \( \gamma \) of the spheroid was identical with the half-length \( l \) and that the spheroid was inclined by \( \theta \) from the vertical axis. Unfortunately, both of them do not reflect the true geometry. The half-length \( l \) should be the distance between the vortex center and the top point of the spheroid, at which it reaches the horizontal plane \( z = Z + zh \). Similarly, the angle \( \theta = \cos^{-1}(zh/l) \) does not coincide with the inclination angle \( \hat{\theta} \) of the axis of symmetry of the spheroid (see Figure 2).

Careful geometrical reconsideration yields the correct relations:

\[ \hat{\varepsilon}^2 = \frac{\tan(\hat{\theta}-\theta)}{\tan \hat{\theta}}. \] (31)

The relation between \( \theta \) and \( \hat{\theta} \) is rather intricate:

\[ \tan^3 \hat{\theta} + \left[ (1 - \hat{\varepsilon_0}) \cot \theta - 2 \tan \theta \right] \tan^3 \hat{\theta} + (\tan^2 \theta - 2) \tan \hat{\theta} + \tan \theta = 0. \] (32)

![Figure 2 Variables specifying a spheroidal vortex.](image)

The exact solution of this algebraic equation of cubic order is provided by Cardan's formula. We select the real solution that is larger than \( \theta \). There still remains another subtle point. Even using the geometrically correct relation (31) with the solution of (32), the
self-energy (28) (in which \( l \) should be replaced by \( \gamma \) ) and the self-rotation rate (29) are inconsistent. The correct self-rotation rate can be obtained by the relation

\[
\dot{\Omega}_s = -\frac{\partial H_s}{\partial S},
\]  

(33)

with the correct geometrical relation (see also Miyazaki et al. (2001))

\[
S = \frac{\Gamma zh^2}{10} \left( \frac{1 - \varepsilon^2}{\cos^2 \theta \sin^2 \theta} \right)^2 \left( \sin^2 \theta \right)^2 \left( \cos^2 \theta + \varepsilon^2 \sin^2 \theta \right)^2
\]

\[
= \frac{\Gamma zh^2}{10} \left( 1 - \varepsilon_0^4 \right) ^{1/3} \left( \varepsilon_0^2 \right) ^{2/3} \left( \varepsilon_0^2 \right) ^{4/3} \left( 1 - \varepsilon \right)
\]

(34)

After some algebra, we notice a factor should be inserted:

\[
\dot{\Omega}_s = \frac{\Omega_s}{1 + \varepsilon^2 - \varepsilon_0^2 \varepsilon^2 \left( \varepsilon^2 \right)^{2/3}}
\]

(35)

The reason why such a delicate correction factor appears is not clear, at present. A systematic reduction of variables from the full ellipsoidal model (3N degrees of freedom) to the spheroidal model (2N degrees of freedom) may provide the answer, which is left for future work. Anyway we have derived the refined spheroidal vortex model, whose performance is checked in the following sections.

3. Counter-rotating Vortex Pair

We investigate the interaction between two counter-rotating vortices of the same strength \( (\Gamma_1 = \Gamma_2 = 1) \) and the same shape \( (zh_1 = zh_2 = 1, \varepsilon_0 = \varepsilon_0 = 0.1) \), based on our refined model. We have integrated the equations (2), (3), (6) and (7) numerically, using the package software ‘LSODE’. We start the computations from the initial conditions, \( X_2 = X_1 = a, Z_2 - Z_1 = h \) with \( Y_1 = Y_2 = 0 \). We show the prediction of the moment model in Figure 3. In the region (1), the vortices translate in the positive y-direction with weak precession. They precess largely and translate in the negative y-direction, if they are within the critical distance and \( h/a > \sqrt{2} \) (region (2)). Both vortices are stretched infinitely, when they are placed within the critical distance and the condition \( h/a < \sqrt{2} \) is satisfied (region (3)). The latter two cases are never observed in the CASL-computation of the original quasi-geostrophic equation.
Figure 3 Critical distance of singular behavior: Wire moment model. The horizontal axis denotes the initial horizontal distance $a$ and the vertical axis does the initial vertical distance $h$. The shadowed region is the region of initial overlap of two vortices. (1) Stable translation, (2) Large precession and (3) Infinite stretching.

We suspect that the inaccuracy in evaluating the mutual energy is the main reason of the above discrepancy. We see in Figure 4 that our refined model (with the Gaussian integration method with 2-5 points) predict no singular behavior, as long as two vortices do not overlap initially, i.e., the critical distance falls into the shadowed region, which denotes the initial overlap of two vortices. The model becomes a little robust using odd number points than using even number points, i.e., the center of each vortex should be included in the set of integration points. The refinement on the approximation of the self-energy also makes the model robust, but its effect is very weak for the case considered here ($\varepsilon_{\alpha 1} = \varepsilon_{\alpha 2} = 0.1$), because the vortices are very slender. The main improvement is provided by the Gaussian integration of the mutual energy. We have investigated the interaction between two fatter vortices with $\varepsilon_{\alpha 1} = \varepsilon_{\alpha 2} = 0.3162$. They do not show singular behavior at all.

Figure 4 Critical distance of singular behavior: Gaussian integration. See the caption of Figure 3 for details. Singular behavior is never observed outside the shadowed region.

4. Merger of Co-rotating Vortices

In this section, we study the interaction between two co-rotating wire-vortices with $\Gamma_1 = \Gamma_2 = 1$, $zh_1 = zh_2 = 1$ and $\tan \theta_1 = \tan \theta_2 = 0.0141$. The vortices are tilted slightly, because initial asymmetry is necessary for chaotic interaction. We use the two-, three- and
five-points Gaussian integration and the slenderness parameters \( \varepsilon_{0.1,0.2} \) are taken to be 0.3162 (Miyazaki et al. 2000).

This value corresponds to the case studied in Miyazaki et al. (2002) in which the predictions of the ellipsoidal moment model were compared with the results of the direct numerical simulation (CASL). Here, we assess the influence of both refinements, re-comparing the predictions of the refined wire model with the results provided by the CASL-computations.

As in the previous section, we have integrated the equations (2), (3), (6) and (7) numerically, starting from various initial locations, i.e., changing \( X_2 - X_1 = a, \ Z_2 - Z_1 = h \) with \( Y_1 = Y_2 = 0 \). We think that two vortices merge when they overlap vertically \( z_1 + z_2 \geq |Z_1 - Z_2| \) and horizontally \( \frac{z_1 \tan \theta_1 + z_2 \tan \theta_2}{\sqrt{(X_1 - X_2)^2 + (Y_1 - Y_2)^2}} \).

The following results are not sensitive to the details of merger-criteria, because the inclination angles increase abruptly as soon as chaotic motions set in.

Figure 5 shows the critical distance of merger (various lines) determined by the refined wire model. The results due to the CASL-computations are essentially carried over from Figure 8 of Miyazaki et al. (2002). Three types of interaction are found in the CASL-simulations, i.e., no-merger (stable rotation), merger and intermediate cases. The crosses show the stable rotation around the vorticity-center. The open circles denote the merger, whereas the open triangles do the cases of intermediate behavior. The vortices merge once, then break into two pieces that are different from the initial vortices. We have performed the CASL-computations with finer mesh points (128^3) at several points near the boundaries to make the classification more reliable. The solid line denotes the critical distance predicted by the ellipsoidal moment model. The correspondence to the CASL-results is not bad as a whole. The ellipsoidal moment model, however, overestimates the interaction between slightly off-set vortices \( h<0.8 \), and it underestimates the interaction between largely off-set vortices \( h>1 \).

![Figure 5](image)

**Figure 5** Critical distance of merger of two co-rotating vortices with \( \varepsilon_{0.1,0.2} = 0.3162 \). Green line: moment model, The thinnest line: two-points approximation, Thin blue line: three-points approximation, Blue line: five-points approximation. Circle: full merger (CASL), Triangle: touch and go (CASL), Cross: stable rotation (CASL)

We draw the critical distance due to the refined model with the two- (blue line), three- (thin blue line) and five- (the thinnest line) points approximation of the mutual energy integral.
We also use the refined expression of the self-energy (28) with the refined self-rotation rate (35). The model predictions are improved considerably, even if we adapt the two-points integration. The lines due to the three- and five-points approximation are almost indistinguishable.

The spheroidal vortex model with 2N degrees of freedom is far simpler than the ellipsoidal moment model with 3N degrees of freedom. The refined wire model is satisfactory for slender vortices. It will be of interest to check whether it can model fatter vortices, comparing the results both by the CASL-computation and by the ‘refined’ ellipsoidal model. The ellipsoidal vortex model newly developed by Dritschel et al. (2004) is certainly a good candidate, but it is rather complicated using non-canonical variables. Further simplification to a canonical form will be helpful for later use.

5. Summary

We have refined the vortex model in two ways. First, the Gaussian integration method is introduced in the evaluation of the mutual energy integral, and then the self-energy is approximated by that of a prolate spheroidal vortex. The former refinement improves the accuracy of the mutual energy computation drastically and the refined spheroidal model does not suffer from the singular behavior in the interaction between two counter-rotating vortices. The latter improvement extends the applicability of the model for fatter vortices. In fact, the refined model can predict the critical merger distance of co-rotating vortices much better than the ellipsoidal moment model.

Acknowledgements

The CASL-computations were performed on the super-computer (NEC SX-6) of the Center for Global Environmental Research (CGER), National Institute for Environmental Studies (NIES).

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Development of an Urban Meteorological Numerical Model in Cartesian Coordinate

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Abstract
We intend to develop a new urban street atmospheric numerical model in Cartesian coordinate, which is expected to treat any complex object (buildings, etc.) explicitly in urban street with a finer resolution. In this work, the finite volume method (FVM) in conjunction with the Semi-Implicit Method for Pressure-Linked Equation Revised (SIMPLER) algorithms is used for calculations on a staggered grid. Abandoning the customary terrain-following normalization, we choose the Cartesian coordinate in which the height is used as the vertical one. Blocking-off method is introduced to handle all of complex objects. For the spatial and temporal discretizations, higher-order upwind convection scheme is employed, and fully time implicit scheme is utilized. As a test, the model has been run on flows over complex building blocks in downtown city, Tokyo, by Large Eddy Simulation (LES). Result of simulations is present which shows the potential of our proposed approaches for the new urban scale model development.

Keywords: Urban flow modeling, Cartesian coordinate, Time implicit scheme, Higher-order upwind convection scheme, Finite volume method, Complex object, Blocking-off method

1. Introduction

With the continuing increase in computational resources, the meso-scale models are being run at successively higher resolutions. It is reasonable to expect that a goal of running a regional model at a very high horizontal resolution may be attainable in the near future, and the topography and objects (e.g., buildings in urban city) could be more accurately represented. In such a situation, it seems natural to search for an alternative that will be better suited to handle the step topography and complex objects on the surface for the high-resolution models currently used as well as a future model expected to run with a finer resolution.

Since terrain-following vertical coordinate (sigma) system (Phillips, 1957; Gal-Chen and Somerville, 1975) has been used extensively to accommodate orography in models for atmospheric flows, most of existing community meso-scale atmospheric numerical models in the world are using the terrain-following coordinate as the vertical coordinate. However, a problem that has received attention rather early in the development of sigma system primitive equation models is the noncancellation errors in the two terms of the gradient force in the momentum equation (Smagorinsky et al., 1967). The two terms on the expression of the pressure gradient force have comparable magnitudes with opposite sign over steep topography and thus their sum may be subject to large errors. Mesinger and Janjic (1985), among others (Sundqvist, 1976), have found that errors in computing the horizontal pressure gradient force in models using a sigma coordinate can be substantial in the vicinity of steep topography. To minimize this error, a step-mountain vertical coordinate, the so-called eta
coordinate, is implemented in the National Centers for Environmental Prediction (NCEP) Meso Eta Model (Meesinger et al., 1988) in which the topography is represented as discrete steps (step mountain). However, the step-mountain representation can cause spurious perturbations at step corners and its accuracy may depend strongly on the horizontal scale of the terrain and the resolution of the actual terrain by the vertical grid (Gallus and Klamp, 2000). Recently, representation of topography, i.e., the “shaved cell” approach, and the related numerical schemes for the equations of geophysical flows in ocean and atmosphere models in which the height is used as vertical coordinate, have been proposed formulated on the finite volume method (Adcroft et al., 1997; Marshall et al., 1997; Bonaventura, 2000).

In this report, we present the advanced numerical method based on finite volume discretization, i.e., blocking-off method for handling complex geometry, which has been incorporated into coding of a robust, efficient and accurate dynamical core for a new atmospheric meso/urban scale numerical model which is expected to suitably treat the steep topography and complex objects (e.g., buildings in urban city) with a finer resolution for the high-resolution flow simulations.

2. Model Descriptions

2.1 Governing Equations

Three-dimensional, unsteady Navier-Stokes equations for viscous compressible Newtonian fluid are used;

Momentum equations:
\[
\begin{align*}
\frac{\partial \rho u}{\partial t} + \frac{\partial \rho uu}{\partial x} + \frac{\partial \rho vv}{\partial y} + \frac{\partial \rho w w}{\partial z} &= -\frac{\partial \rho}{\partial x} + \rho f_v + \mu \frac{\partial^2 u}{\partial x^2} + \mu \frac{\partial^2 u}{\partial y^2} + \mu \frac{\partial^2 u}{\partial z^2} \\
\frac{\partial \rho v}{\partial t} + \frac{\partial \rho vu}{\partial x} + \frac{\partial \rho vv}{\partial y} + \frac{\partial \rho w w}{\partial z} &= -\frac{\partial \rho}{\partial y} - \rho f_u + \mu \frac{\partial^2 v}{\partial x^2} + \mu \frac{\partial^2 v}{\partial y^2} + \mu \frac{\partial^2 v}{\partial z^2} \\
\frac{\partial \rho w}{\partial t} + \frac{\partial \rho w u}{\partial x} + \frac{\partial \rho w v}{\partial y} + \frac{\partial \rho w w}{\partial z} &= -\frac{\partial \rho}{\partial z} - g + \mu \frac{\partial^2 w}{\partial x^2} + \mu \frac{\partial^2 w}{\partial y^2} + \mu \frac{\partial^2 w}{\partial z^2}
\end{align*}
\]

Energy equation:
\[
\frac{\partial \rho T}{\partial t} + \frac{\partial \rho T}{\partial x} + \frac{\partial \rho T}{\partial y} + \frac{\partial \rho T}{\partial z} = \kappa \frac{\partial^2 T}{\partial x^2} + \kappa \frac{\partial^2 T}{\partial y^2} + \kappa \frac{\partial^2 T}{\partial z^2}
\]

Continuity equation:
\[
\frac{\partial \rho}{\partial t} + \frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y} + \frac{\partial \rho w}{\partial z} = 0
\]

Equation of state for ideal gas:
\[
P = \rho RT
\]

2.2 Temporal and Spatial Discretizations

Detail of the numerical scheme on the temporal and spatial discretization used in the model is referred to Sha (2002, 2004).
2.3 Treatment of Irregularly Shaped Objective in Calculation Domain

We describe the manner in which we treat arbitrary geometries by the blocking-off method (Patankar, 1980). This is done by blocking off some of the control volumes of the regular grid so that the remaining inactive control volumes form the desired irregular domain. The arbitrary geometries are approximated by a series of the rectangular grids.

The idea of the blocking-off operation consists of establishing known values of the relevant $\phi$ 's in the inactive control volumes. The desired values can be obtained in the inactive control volumes by setting a large source term in the discretization equations, which denotes a number large enough to make the other terms in the discretization equation negligible. This procedure is easily used to represent irregularly shaped objective in the calculation domain by inserting such internal boundary conditions.

3. Test Result

The model has been run by Large Eddy Simulation (LES) of real turbulent flows within the downtown city, i.e., Ootemati, Tokyo. Figure 1 shows the building blocks and the grids set in the calculation. In Figure 2, we present the wind velocity and temperature near the surface. It illustrates the flow pattern characterized by the block, and the simulation shows a satisfying result. Next work on combining radiation calculation in the model is in progress.
Figure 2  Wind velocity and temperature near the surface.

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Conference Report:

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4. Other Research
The High-Resolution Numerical Model of Heat Island Phenomena

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Abstract

The heat island phenomena have been analyzed in various research fields, and recently these are becoming to be considered as political issues. In the former meso-scale model, cities were indicated as flat, and mesh resolution was rather coarse (in several kilometers). As a result, it was difficult to calculate building effects. This research aims to develop a numerical simulation tool which can deal with multi-scales, from urban districts to regional areas. We attempt to make the meso-scale model coupled with the urban canopy model, called Urban Climate Simulation System (UCSS), which can predict temperatures and velocities in urban areas considering building height and density. This research is planning to verify the new urban canopy model in comparison with Computational Fluid Dynamics (CFD) and the wind tunnel experiment. This year, we installed UCSS and CFD codes into the supercomputer.

Keywords: Heat island, Urban canopy model, UCSS, Wind tunnel experiment

1. Introduction

Heat island phenomena have been analyzed in various research fields, and recently these are becoming to be considered as political issues. For example, the Committee of the Comprehensive Regulation Reform conducted by the Cabinet Office held up the “Promotion of the investigation and research on the mechanism of the heat island phenomena”.

In practice, countermeasures are usually implemented at the building scale, but heat island is a phenomenon observed at the macro scale, which forms as a result of the accumulation of the urban effects. Therefore, it is important to evaluate validity of the countermeasures in the urban scale. In order to clarify the mechanism of the heat island and evaluate countermeasures, it is necessary to get over the scale gap, when performing numerical simulation.

In the architectural environment field, model researches on the urban canopy layer have been seen during last decades. The urban canopy model is developed from the viewpoint of the architectural planning. Accordingly, mechanism of the air conditioner and thermal properties of the architectural components become explicit in the urban canopy model, which were buried under the sub grid in the roughness model. As a result, it is expected to consider the transportation of heat and momentum more accurately in the urban canopy model. However, case studies by the urban canopy model are few, and verifications of the model have not been performed sufficiently.

The Building Research Institute is developing the Urban Climate Simulation System (UCSS) as a tool to simulate heat island phenomenon. In this research, we are planning to verify the urban canopy model adopted in the UCSS, using the Reynolds Average Numerical Simulation (RANS) and the Large Eddy Simulation (LES) analysis by the supercomputer in the NIES.
2. Research Objective

Table 1 shows a comparison of the various models of urban structure. The roughness model is mainly used for wide areas, hundreds of kilometers on all sides, but it is difficult to resolve the details of urban structures. On the other hand, CFD can analyze the changes of temperature and ventilation, when building location is modified in the plan. However, CFD takes enormous time for calculation. If we regard a lot of buildings as one unit by space averaging method, the whole city can be analyzed with the resolution of urban structures. However, verification of the urban canopy model is not enough, so it is required to perform a large number of case studies in order to fix model parameters.

<table>
<thead>
<tr>
<th>Model type</th>
<th>Resolution</th>
<th>Computational cost</th>
<th>Representation of buildings</th>
<th>Main input data</th>
</tr>
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<tbody>
<tr>
<td>Roughness model</td>
<td>Coarse</td>
<td>Low</td>
<td>Flat surface with roughness</td>
<td>· Physical parameters of land use (roughness, albedo, evaporation rate, thermal conductivity)</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>· Anthropogenic heat</td>
</tr>
<tr>
<td>Urban canopy model</td>
<td>Medium</td>
<td>Medium</td>
<td>Bulk density and average height of buildings</td>
<td>· Settings of buildings (building ratio, building height, air conditioning system, etc.)</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>· Surface covering properties (albedo, evaporation rate, thermal conductivity)</td>
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<tr>
<td>CFD</td>
<td>Fine</td>
<td>High</td>
<td>Building shapes and arrangements</td>
<td>· Shape, arrangement of buildings, roads and trees</td>
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<td></td>
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<td>· Surface covering properties (albedo, evaporation rate, thermal conductivity)</td>
</tr>
</tbody>
</table>

Figure 1  Research outline.
Figure 1 shows the outline of this research. This research aims to develop a numerical simulation tool which can deal with multi-scale, from urban districts to regional areas. We attempt to make the meso-scale model coupled with the urban canopy model called UCSS, which can predict temperatures and velocities in urban areas considering building height and density.

Model parameters, such as drag coefficient, are fixed by the data of CFD analysis and wind tunnel experiment. After that, meso-scale simulations are performed by UCSS including the urban canopy model, and the numerical results are compared with AMeDAS data.

3. Development of the UCSS

In order to evaluate the urban planning from the point of view of urban environment, it is desired to make a simulation model capable of considering urban planning factors. The urban GIS data is extremely useful in operating this work, so it is utilized as input data of UCSS. Figure 2 shows the outline of the solver part of UCSS (Ashie and Vu, 2004).
\begin{align*}
\frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_i} &= 0 \\
\text{Velocity equation} & \\
\frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_j} + \frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_j} &= \frac{1}{G} \frac{\partial (\vec{u})}{\partial x_j} + \frac{1}{G} \frac{\partial (\vec{u})}{\partial x_j} - F_i
\end{align*}

\begin{align*}
\text{Momentum equation} & \\
\frac{\partial (\vec{u})}{\partial t} + \frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_j} &= \frac{1}{\rho} \frac{\partial (\vec{P})}{\partial x_j} + \frac{1}{G} \frac{\partial (\vec{u})}{\partial x_j} \left[ GV \left( \frac{\partial (\vec{u})}{\partial x_j} + \frac{\partial (\vec{u})}{\partial x_j} \right) \right] - F_i
\end{align*}

\begin{align*}
\text{Transport equation for } \varepsilon & \\
\frac{\partial \varepsilon}{\partial t} + \frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_j} &= \frac{1}{\rho} \frac{\partial (\vec{P})}{\partial x_j} + \frac{1}{G} \frac{\partial (\vec{u})}{\partial x_j} \left[ GV \left( \frac{\partial (\vec{u})}{\partial x_j} + \frac{\partial (\vec{u})}{\partial x_j} \right) \right] - F_i
\end{align*}

\begin{align*}
\text{Transport equation for } \varepsilon & \\
\frac{\partial \varepsilon}{\partial t} + \frac{1}{G} \frac{\partial G(\vec{u})}{\partial x_j} &= \frac{1}{\rho} \frac{\partial (\vec{P})}{\partial x_j} + \frac{1}{G} \frac{\partial (\vec{u})}{\partial x_j} \left[ GV \left( \frac{\partial (\vec{u})}{\partial x_j} + \frac{\partial (\vec{u})}{\partial x_j} \right) \right] - F_i
\end{align*}

\begin{align*}
\text{here,} & \\
F_i &= \alpha_c C_{l} \left( \frac{\langle \varepsilon \rangle^2}{\langle \varepsilon \rangle} \right)^{1/2} \\
P_k &= \nu \left( \frac{\langle \vec{u}^2 \rangle}{\langle \varepsilon \rangle} \right) \\
F_k &= \frac{\varepsilon}{L} \\
F_c &= \frac{\varepsilon^{3/2}}{L}
\end{align*}

\begin{align*}
u_i \text{ : wind velocity in } x_i \text{ direction, } t \text{ : time, } \rho \text{ : air density, } P \text{ : pressure function, } \\
\varepsilon \text{ : turbulent energy, } \varepsilon \text{ : viscosity dispersion rate, } v_i \text{ : eddy viscosity coefficient, } \\
C_{l} \text{ : canopy model coefficient, } \alpha_x \text{ : area of body surface seen from the } \\
\text{windward in } x_i \text{ direction per fluid volume, } L \text{ : length scale } \\
C_{\mu} = 0.09, \sigma_k = 1.00, \sigma_\varepsilon = 1.30, C_{l} = 1.44, C_{x} = 1.92
\end{align*}

Figure 3 Fundamental equations of UCSS.

Figure 4 shows the relationship between flux Richardson number $R_f$ and model parameter $C_{\mu}$. Relationship between $R_f$ and Prandtl number $P_n$ is shown in Figure 5. The variety of $P_n$ of this model corresponds relatively well to the results obtained by the outdoor measurement and the wind tunnel (Vu et al., 2002).
4. Research Method and Results

In this year, the code of the standard $k$-$\varepsilon$ model was transplanted on SX-6, and accuracy was verified for the flow field around a single block. The vectorization and parallelizing tuning of UCSS were accomplished, and computational speed was improved by 6.9 times. Furthermore, the transplantation of the LES code was carried out, and the improvement in the rate of vectorization, parallelizing tuning and sauce modifications were performed. Also, the computational speed was improved by 10.5 times.

The calculation condition and result of the verification of the $k$-$\varepsilon$ model are as follows.

4.1 Calculation Condition

4.1.1 Calculation Domain

The calculation domain is $21B(x) \times 13.75B(y) \times 11.25B(z)$ (here, B: building width) and the number of grid points is $69(x) \times 45(y) \times 39(z) = 105,300$.

4.1.2 Boundary Conditions

1. Approach flow
   $V=W=0$. $U$ and $k$ were set as the wind tunnel experiment (Yan et al., 1998). $U$ at the height of $Z=0.0625H$ and $Z=H$ (here, $H$: Building height, $=2B$) is 2.94 m/s and 4.49 m/s, respectively (Tominaga et al., 1999).

2. Outflow boundary
   Neumann condition

3. Block and floor surface boundary
   Generalized logarithm low

4.1.3 Integration of Time

The Euler implicit scheme was used for time integration, and computational time was integrated for about 0.605 (1000 iteration step).

4.2 Calculation Result

The cross section of the wind vector distribution around the block is shown in Figure 6 and Figure 7. The reattachment length of the rooftop and building back were compared with the data of references as shown in Figure 8. Similar to the numerical result (Tominaga et al., 1999), the separated flows did not reattach on the rooftop. The reproducibility of reattachment of flows in the downstream of the building was improved compared to that in a previous report (Tominaga et al., 1999).

![Figure 6 Wind vector (vertical section).](image1)

![Figure 7 Wind vector (horizontal).](image2)
<table>
<thead>
<tr>
<th></th>
<th>$X_R$</th>
<th>$X_F$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Exp. Result (Yan et al., 1998)</td>
<td>0.52B</td>
<td>1.42B</td>
</tr>
<tr>
<td>Numerical Cal. Result (Tominaga et al., 1999)</td>
<td>None</td>
<td>2.27B</td>
</tr>
<tr>
<td>This study</td>
<td>None</td>
<td>1.50B</td>
</tr>
</tbody>
</table>

$X_R$ : Reattachment length on the roof  
$X_F$ : Reattachment length of the recirculation in the building backward.

**Figure 8** Reattachment length on the roof and the recirculation in the building backward.

**5. Conclusion**

The works performed this year are as follows.
1) The code of the standard k-ε model was transplanted on SX-6, and accuracy was verified for the flow field around a single block.
2) The vectorization and parallelizing tuning of UCSS were accomplished, and the computational speed was improved by 6.9 times.
3) The transplant of the LES code was also carried out. The vectorization, parallelizing tuning and sauce correction were performed, and improvement in the computational speed by 10.5 times was attained.

Model parameters, such as drag coefficient will be fixed by the data of CFD analysis and wind tunnel experiment. Next, meso-scale simulation will be performed by UCSS including the urban canopy model. We also intend to compare numerical simulation results with AMeDAS.

**References**


**Conference Report**

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Drag and Lift Forces Acting on a Spherical Droplet in a Homogeneous Shear Flow

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Abstract
Drag and lift forces acting on a spherical droplet in a homogeneous linear shear flow were numerically studied by means of a three-dimensional direct numerical simulation based on a marker and cell (MAC) method. The effects of fluid shear rate and particle Reynolds number on drag and lift forces acting on a spherical droplet were investigated. The results show that the drag force acting on a spherical droplet in a linear shear flow increases with shear. The lift force acts on a droplet from high-speed side to low-speed side in the linear shear flow with high particle Reynolds number of \( Re_p > 100 \). The direction of the lift force is different between low and high particle Reynolds number flows. The behavior of the lift force on a droplet is quite similar to that on a rigid sphere.

Keywords: Drag and lift forces, Flow structure outside and inside a droplet, Numerical simulation

1. Introduction

The dispersion phenomena of water droplets are often seen in atmospheric boundary layer with rainfall or near stormy breaking waves in oceans and lakes. It is of importance to investigate the effects of mean shear on fluid forces on a water droplet in making reliable models for expressing the effects of rainfall and droplets generated by breaking waves on the mass transfer across the air-sea interface.

When a rigid sphere is moving in a shear flow, a transverse force referred to as lift force is exerted. The inviscid and low Reynolds number theories predicted the lift force acts from lower fluid velocity side to higher fluid velocity side. Kurose & Komori (1999) first performed a three-dimensional direct numerical simulation around a rigid sphere in the particle Reynolds number range of \( Re_p = 1-500 \), \( Re_p = dU_c/\nu \). Here, \( d \) is the diameter of a sphere, \( U_c \) is the fluid velocity on the streamline through the center of a sphere, and \( \nu \) is the kinematic viscosity. In contrast, they observed that the direction of the lift force acting on a rigid sphere at higher \( Re_p \) is opposite to that predicted by the inviscid and low Reynolds number theories. On the other hand, the studies of the shear lift force on a spherical bubble for high particle Reynolds numbers are limited. Mei & Klausner (1994) derived an expression of lift force on a spherical bubble in a linear shear flow by combining the results for a rigid sphere in a viscous shear flow by Saffman (1965), Dandy & Dwyer (1990) and McLaughlin (1993) with the results for a rigid sphere in an inviscid shear flow by Auton (1987). Legendre & Magnaudet (1998) and Kurose et al. (2001) also estimated the lift force acting on an inviscid spherical bubble in a viscous shear flow numerically.

However, in the above studies, only the ambient flow field outside a sphere is computed. To accurately estimate the fluid force acting on a spherical droplet, it is necessary to compute
flow fields both outside and inside a spherical droplet. In a uniform unsheared creeping flow \((Re_p<1)\), flow fields outside and inside a droplet have been analyzed, but the studies for the moderate particle Reynolds numbers \((Re_p>1)\) are limited (Leclair et. al., 1972; Chen, 2001). Moreover, these studies used unrealistic assumption that the flow fields are axisymmetric. This suggests that the lift force acting on a spherical droplet in a shear flow has not been precisely estimated. And previous numerical estimations of the drag force on a spherical droplet may not be accurate, because when vortex shedding appears in the high Reynolds number range, flow structure is not axisymmetric.

The purpose of this study is to investigate the effects of fluid shear on lift force acting on a spherical droplet in a viscous linear shear flow with a high particle Reynolds number by means of a three-dimensional numerical simulation.

2. Numerical Simulation

The flow geometry and coordinate system for the present direct numerical simulations are shown in Figure 1. The ambient flow outside a spherical droplet was a linear shear flow. The droplet was assumed to be so small that deformation can be neglected.

![Figure 1 Coordinate system for a spherical droplet.](image)

The three-dimensional Navier-Stokes (NS) equations in the cylindrical coordinates \((x, r, \theta)\) are given by

\[
\begin{align*}
\frac{\partial U}{\partial t} + (V \cdot \nabla) U &= -\frac{\partial p}{\partial x} + \frac{1}{Re_{p,k}} \nabla^2 U, \\
\frac{\partial V}{\partial t} + (V \cdot \nabla) V &= -\frac{\partial p}{\partial r} + \frac{1}{Re_{p,k}} \left( \nabla^2 V - \frac{V}{r^2} - \frac{2}{r^2} \frac{\partial W}{\partial \theta} \right), \\
\frac{\partial W}{\partial t} + (V \cdot \nabla) W &= -\frac{V W}{r} - \frac{1}{r} \frac{\partial p}{\partial \theta} + \frac{1}{Re_{p,k}} \left( \nabla^2 W - \frac{W}{r^2} + \frac{2}{r^2} \frac{\partial V}{\partial \theta} \right).
\end{align*}
\]

(1)

The particle Reynolds number \(Re_p (= \rho \cdot dU/\mu)\) is based on the mean velocity of the ambient
fluid on the stream through the center of the sphere $U_c$. $\rho$ is the density of the fluid, $d$ is the diameter of a droplet and $\mu$ is the viscosity. The subscript $k$ indicates $o$ for the ambient flow outside a droplet and $i$ for the flow inside a droplet, respectively. The particle Reynolds number outside a droplet $Re_{p,o}$ and the particle Reynolds number inside a droplet $Re_{p,i}$ are not independent and they are related by

$$Re_{p,i} = \frac{\rho_i \mu_i}{\rho_o \mu_o} Re_{p,o}.$$

(2)

The fluids outside and inside a droplet were air and water, respectively. Physical properties of fluids are shown in Table 1.

<table>
<thead>
<tr>
<th>Fluid properties</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>$\mu_o$ [kg/ms]</td>
<td>$18.2 \times 10^{-6}$</td>
</tr>
<tr>
<td>$\mu_i$ [kg/ms]</td>
<td>$890.9 \times 10^{-6}$</td>
</tr>
<tr>
<td>$\rho_o$ [kg/m$^3$]</td>
<td>$1.185$</td>
</tr>
<tr>
<td>$\rho_i$ [kg/m$^3$]</td>
<td>$997.0$</td>
</tr>
</tbody>
</table>

The NS equations were directly solved using a finite difference scheme based on the marker and cell (MAC) method. Numerical procedure used here was essentially the same as that used in Hanazaki (1988), Kurose & Komori (1999) and Kurose et al. (2001). Numerical grids outside and inside a spherical droplet are shown in Figure 2 (a) and 2 (b). The grid points outside a spherical droplet used in this study were $35 \times 61 \times 48$ in $\eta, \xi$ and $\theta$ directions. And the grid points inside a droplet were $35 \times 31 \times 48$. The boundary condition of velocity upstream of a spherical droplet was given in a dimensionless form by

$$U = 1 + \alpha y.$$

(3)

Here, $\alpha$ is the dimensionless fluid shear rate of the mean flow. The boundary conditions on the surface of a droplet were given by no flow condition across the interface and the continuities of tangential velocity and stress:

$$v_{n,o} = v_{n,i} = 0$$
$$v_{\theta,o} = v_{\theta,i}$$
$$v_{\phi,o} = v_{\phi,i}$$
$$\tau_{n\theta,o} = \tau_{n\theta,i}$$
$$\tau_{n\phi,o} = \tau_{n\phi,i}$$

(4)

The $\tau$ denotes the viscous stress. Subscript $n$ denotes the normal direction to the surface of a droplet. The subscripts of $\theta$ and $\phi$ show the tangential directions to the surface of a droplet.

The fluid force acting on a spherical droplet is computed by integrating the pressure and
viscous stresses over the surface of the spherical droplet:

$$ F = \int_S -p e_n + \tau_{ni} e_i \, dS $$ \hspace{1cm} (5)

The first term on right-hand side is the pressure term, while the next term is the viscous term. A dimensionless force coefficient $C$ is defined as

$$ C = \frac{F}{\frac{1}{2} \pi \rho U_e^2 \left( \frac{d}{2} \right)^2}. $$ \hspace{1cm} (6)

The component of $C$ along the streamwise direction is the drag coefficient $C_D$. Due to the three-dimensional nature of the flow, a lift force normal to the streamwise direction is generated, and the lift coefficient is denoted by $C_L$.

The computations were performed for particle Reynolds numbers of $Re_p=1, 5, 10, 50, 100$ and $300$ and for fluid shear rates of $\alpha=0.0, 0.1, 0.2, 0.3$ and $0.4$.

![Figure 2 Schematic diagram of computational domain: (a) outside a spherical droplet. (b) inside a spherical droplet.](image)

3. Results and Discussion

3.1 Drag in a Linear Shear Flow

Figure 3 (a) and 3 (b) show the velocity fields and streamlines outside and inside a spherical droplet in a uniform unsheared flow ($Re_p=300, \alpha=0.0$) on the center-line ($z=0$). It is found that vortex shedding appears in the ambient air flow. Both air and water flow fields are not axisymmetric, and double-vortex motions inside a droplet (Leclair et al., 1972; Chen, 2001) are not found.

Figure 4 shows the variations of drag coefficient $C_D$ on a spherical droplet in a uniform unsheared flow against the particle Reynolds number $Re_p$. The computed $C_D$ on a spherical droplet in Figure 4 is compared with the empirical expressions for a spherical droplet by Beard & Pruppacher (1969) and Gunn & Kinzer (1949). The present results for a spherical droplet...
droplet are in good agreement with their expressions. Figure 5 shows the effect of fluid shear rate $\alpha$ on drag coefficient $C_D$. The ratio of $C_D$ to the drag coefficient in a uniform unsheared flow $C_{D0}$ is plotted as a function of $Re_p$ against the dimensionless shear rate $\alpha$ in Figure 5. $C_D$ increases with increasing $\alpha$ for a fixed value of $Re_p$, and dependence of $C_D$ on $\alpha$ is more obvious for higher $Re_p$.

![Schematic diagrams of velocity fields and stream lines at $Re_p = 300$ and $\alpha = 0.0$: (a) outside a spherical droplet; (b) inside a spherical droplet.](image)

Figure 3  Schematic diagrams of velocity fields and stream lines at $Re_p = 300$ and $\alpha = 0.0$: (a) outside a spherical droplet; (b) inside a spherical droplet.

![Drag coefficient $C_D$ on a spherical droplet in a uniform unsheared flow versus the particle Reynolds number $Re_p$.](image)

Figure 4  Drag coefficient $C_D$ on a spherical droplet in a uniform unsheared flow versus the particle Reynolds number $Re_p$. 
3.2 Lift in a Linear Shear Flow

Figure 6 shows the variations of the lift coefficient $C_L$ on a spherical droplet and in a linear shear flow $\alpha = 0.0, 0.1, 0.2, 0.3$ and $0.4$ against the particle Reynolds number $Re_p$. The lift force in a uniform unsheared flow ($\alpha = 0.0$) does not appear ($C_L = 0$). The value of $C_L$ on a spherical droplet in a linear shear flow rapidly decreases with increasing $Re_p$ in the moderate particle Reynolds number range of $Re_p < 10$. In the high particle Reynolds number range of $Re_p \geq 100$, $C_L$ on a spherical droplet shows negative values. The effects of the fluid shear $\alpha$ on $C_L$ increase with increasing $\alpha$. The behavior of the lift force on a droplet is quite similar to that on a rigid sphere (Kurose & Komori, 1999).
4. Conclusion

A three-dimensional direct numerical simulation was done for a linear shear flow outside and inside a spherical droplet with the high particle Reynolds numbers, and the effects of fluid shear on the drag and lift forces on a spherical droplet were investigated.

The drag coefficient on a spherical droplet increases with increasing shear rate for a fixed particle Reynolds number, and dependence of the drag coefficient on shear rate is more obvious for higher particle Reynolds numbers. The lift force on a droplet decreases with increasing the particle Reynolds number, and it acts from the high-speed side to the low-speed side in the ambient linear shear flow with the high particle Reynolds number of $Re_p \approx 100$.

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Original Papers and Reviews:

Conference Reports:
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Use of Surface Erosion Model to Estimate Sediment in Jialingjiang Catchment Upstream of Changjiang River, China

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Abstract

The Three Gorges Dam is expected to tame lethal floods while generating jobs, commerce, and a ninth of China's electricity. As the vast amount of river sediment silts up the lake behind the dam wall, the dam's capacity for hydropower generation and flood control will be diminished, and the decrease of sediments, including nutrients, may alter river morphology and habitats and damage ecosystems. Since the Jialingjiang catchment is one of the main sources of sediment in the Changjiang catchment, knowledge of sediment runoff in this catchment is considered to be indispensable for the purpose of controlling the sediment inflow to the Three Gorges Dam. For this purpose, it is necessary to specify which types of sediment erosion model can be applied to the Jialingjiang catchment. We tested three types of surface erosion (sediment yield) model for 1987, and examined their features, applicability, and limitations.

Keywords: sediment yield, surface erosion, unit stream power, USLE, riverbank erosion

1. Introduction

The Three Gorges Dam will create a 600-km-long lake, into which $5 \times 10^8$ t of sediment and $1 \times 10^9$ t of domestic and industrial wastes will flow each year. In 1987, the upper Changjiang (Yangtze River) tributaries delivered about $0.58 \times 10^9$ t of sediment downstream, and the Three Gorges reach added $0.27 \times 10^9$ t, totaling about $0.85 \times 10^9$ t at the Yichang hydro-station. Approximately 35% of the sediment load came from the Jialingjiang catchment, the biggest sediment load in the upper Changjiang.

The Jialingjiang, at 1120 km long and with a catchment area of 160,000 km$^2$, is one of the main tributaries of the Changjiang, joining the main river at Chongqing, upstream of the Three Gorges Dam (Figure 1). The Jialingjiang river has two main tributaries, the Qujiang at 687 km long to the east, and the Fujiang at 658 km long to the west, both of which join at Hechuang (Figure 2). The climate of the Jialingjiang catchment is close to subtropical and is monsoonal, and the average annual temperature is 16 to 18 $^\circ$C. Precipitation averages 1000 mm a year. The rainy season corresponds to summer and fall, and 70% of the annual precipitation falls from April to October. The topographic characteristics are quite different between the upper and lower regions. Mountains cover the former, and a vast flat area forming the Sichuan Basin covers the latter. As indicated by the digital land cover map produced from Landsat TM data from the early 1990s (Liu, 1996), 53% of the total catchment is steeper than a gradient of 1/100, and agriculture is well developed in both the flat and mountainous regions. Paddy fields and farmlands cover 13% and 39%, respectively, of the total area.
The dominant soil type in the Jialingjiang catchment, as indicated by the China soil map developed by the Nanking Institute of Soil Science (1980) is yellow-brown earth, and covers most of the upper region of the Qujiang. In the upper regions of the Jialingjiang mainstream and the Fujiang, the dominant soil type is burozem (Haplic Luvisol), followed by cinnamon soils. In the lower regions, purplish soil (Eutric Regosol) is dominant and covers around 35% of the total catchment. The China soil texture map shows that loam covers over 99% of the catchment. In particular, sandy loam (>20% sand; particle size, 0.05–1.0 mm) covers around 50% of the catchment, followed by silt containing sand to less 20% and silt to over 40% and sandy silt containing sand to over 20% and silt to over 40%.
When we estimate the sediment yield and transport volume, we have to consider the geophysical and hydrological diversities of the Jialingjiang catchment.

The sediment erosion rate caused by rainfall runoff consists of four sub-mechanisms: pick-up due to rainfall impact, transport due to rainfall impact, pick-up due to overland flow, and transport due to overland flow. Therefore, the sediment erosion rate equation (sediment discharge equation) should be investigated from the viewpoint of these sub-processes.

Generally the Universal Soil Loss Equation (USLE), developed by Wischmeier & Smith (1958) is widely used for a variety of purposes in studies of soil erosion. However, it is pointed out that the USLE is not accurate at predicting soil loss on the lower parts of slopes or under short-term unsteady runoff. To develop a soil erosion model that can be used in such conditions, it is necessary to modify and develop the USLE. The soil erosion rate models calculate the transport rate, $q_B$ (unit volume per unit width and time), of eroded sand as $q_B = \Xi \cdot \Pi \cdot \Gamma \cdot q_{90}$, in which $\Xi$ = vegetation cover factor, $\Pi$ = management factor, $\Gamma$ = support practice factor, and $q_{90}$ = transport rate on bare land without support or management.

However, the USLE focuses more on pick-up due to rainfall impact than on transport processes, which are modeled empirically and cannot be applied to general conditions.

Once an appropriate sediment discharge equation is established, we can estimate sediment yields and transport rate in a catchment. Hence, it is important to establish a model of sediment transport caused by rainfall runoff. For instance, Agricultural Non-Point Source pollution model (AGNPS), Soil & Water Assessment Tool (SWAT), and Hydrological Simulation Program—FORTRAN (HSPF) use the USLE with modifications, while Aerial Non-point Source Watershed Environment Response Simulation (ANSWERS) and Chemicals, Runoff and Erosion from Agricultural Management Systems (CREAMS) combine the USLE with a sediment transport model in river fluvial phenomena.

The USLE and transport model involved in detachment due to rainfall impact and transport due to flow for support of the physical background. However, the transport model derived from fluvial hydraulics is applicable to the mild-slope gradients, but not to the Jialingjiang catchment, because 53% of its total area is steeper than 1/100.

Sediment runoff in monsoon Asia is a strongly unsteady, discrete, intermittent, and non-equilibrium phenomenon caused by complex stochastic agents, so it is not easy to establish a general sediment yield model. For these reasons, we tried to select a sediment erosion and yield model that can model the active sediment runoff from a steep slope. Since slopes, riverbanks, and flood plains are thought to be the main sediment sources, the key is to select erosion models that incorporate these features.

Although the previously proposed models have several disadvantages, they have the advantage of being easy to apply, because they focus not on unsteady, discrete sediment erosion but on sediment erosion that is uniquely determined by a driving agent (e.g., rainfall intensity, slope gradient, overland flow discharge, stream power). Hence, this study focuses on development and applicability of surface erosion models that are easy to apply. We selected three typical types of sediment erosion and yield model:

(i) USLE-type model modified for sediment yield in Japan,
(ii) unit stream power type of fluvial sediment transport rate equation,
(iii) pick-up model of sediment in floodplain.

By dividing sediment yield sources into two areas and considering the intermittent transport down the catchment slope to the river, we could easily evaluate the effects of afforestation, vegetation cover, sabo dam (check dam), and hydraulic structures in each region.
2. Sediment Erosion Model of Catchment Slope

2.1 USLE Applied to Sediment Yield in Japan

Since USLE is designed to estimate annual soil losses from farmland, it is fairly difficult to use it to estimate sediment erosion in Japan caused by short-term heavy rain and long-term intermittent and relatively weakly rain. Hence, several modified USLE-type models have been developed in Japan. Kimoto et. al. (1999) proposed a modified USLE for application to the devastated small basins of southern of China in the following form:

\[ q_b = aK_e(L/19.7)^\beta (\sin \theta)^\alpha f r, \]

in which \( q_b \) = sediment erosion rate per unit width (m³ m⁻¹ h⁻¹), \( K_e \) = sediment erodibility coefficient, \( L \) = slope length (m), \( \theta \) = slope angle (degree), \( f \) = runoff ratio, \( r \) = rainfall intensity (mm hr⁻¹), and \( a, \alpha, \beta \) = empirical parameters. Parameters for bare land were adjusted: \( K_e = 1, \alpha = 1.06, \alpha = 1.2, \beta = 0.7 \).

2.2 Unit Stream Power Type of Sediment Transport Rate

To estimate sediment transport rate (erosion rate) on bare land \( (q_{b0}) \), many researchers have applied a sediment discharge equation to river sedimentation. The equilibrium sediment discharge equation for a river is expressed in the following general form:

\[ q_{b0} = K_0 \Psi(\theta) \tau_{c} \tau_{e} \]

\[ \Psi(\theta) = \frac{\tau_{e}}{\tau_{c}(\theta)} \]

in which \( q_{b0} = q_{b0} / \{(\sigma/\rho - 1)gd\}^{0.5}, q_{b0} \) = sediment discharge per unit width per time, \( K_0 \) = empirical constant for mild slope, \( \sigma = \) density of sand, \( \rho = \) density of water, \( d = \) representative diameter, \( \theta = \) angle of slope, \( \tau = \) non-dimensional tractive force \( (= u^2 / \{(\sigma/\rho - 1)gd\} \), \( u = \) shear velocity), and \( \tau_{c}(\theta) \) and \( \tau_{e} \) = non-dimensional critical tractive forces for a slope with angle \( \theta \).

Generally, the flow structure changes gradually as the relative flow depth changes with slope gradient, producing a change of critical tractive force and the value of \( K_0 \). Since Equations (2) and (3) includes these two effects, it is applicable to a catchment slope whose gradient changes greatly from mountainous regions to flat areas.

In this study we used the HSPF component of the Stanford Watershed Model (Crawford & Linsley, 1966) as rainfall runoff model. General outputs of the catchment runoff model are overland flow discharge, interflow discharge, and groundwater discharge according to land-use. For this reason, when we use the sediment discharge equation expressed in terms of bed shear stress, such an expression of Equation (2) should be rewritten as a sediment discharge equation expressed in terms of overland flow discharge. Since the driving force for soil erosion is flow discharge \( (q) \) and slope gradient \( (l) \), the stream power, \( ql \), that represents the work done by flow for a specific weight is an appropriate index that is equivalent to bed shear stress.

The dimensionless bed shear stress \( \tau \) can be converted to the dimensionless power stream \( q/I \), as follows:

\[ q/I = (\tau_*/\phi)^{2/3}, \]

in which \( q_* = q / \{(\sigma/\rho - 1)gd\}^{0.5}, I_* = I / (\sigma/\rho - 1), d = \) representative diameter, \( I = \) slope gradient, \( q = \) unit discharge of overland flow, \( \phi = \) velocity coefficient (ratio of average velocity of flow to shear velocity), which changes sensitively according to the change in slope angle and relative depth of overland flow.
Substitution of Equation (4) into Equation (2) gives the following sediment discharge equation:

$$q_{r*} = K_0 \cdot \frac{|\Psi(\theta)|}{\phi^{2/3 m} \cdot ((q_* I_*)^{2/3} - (q_* I_*)_c^{2/3})^n}.$$  (5)

In sediment erosion caused by rain-drop impact, the critical unit stream power $(q_* I_*)_c$ at which erosion occurs becomes much smaller than in erosion without raindrop impact. Hence, we can assume that $q_* I_* >> (q_* I_*)_c$, and we approximated the above equation as follows:

$$q_{B_0*} = A_0 \cdot \{(q_* I_*) - (q_* I_*)_c\}^m,$$  (6)

$$A_0 = K_0 \cdot \frac{|\Psi(\theta)|}{\phi^{m}},$$  (7)

in which $m = 2n/3$. Since sediment erosion on bare land include lots of fine sediments, Equation (6) is thought to express total load approximately. Hence, the empirical constant $K_0$ is approximately 10, and the empirical constant $n$ is usually taken to be $5/2$, i.e. $m = 3/5$ (Brown, 1960; Graf and Acaroglu, 1968). On the other hand, the velocity coefficient $\phi$ is within 1–5 for overland flow on a steep slope (Tsujimoto, 1991), and neglecting the effect of slope gradient gives the value of $K_0$ as approximately 1–10. Then Equation (6) is rewritten:

$$q_{B_0*} = A_0 \cdot \{(q_* I_*) - (q_* I_*)_c\}^{5/2},$$  (8)

in which $(q_* I_*)_c$ = empirical non-dimensional critical stream power, and changes from $(q_* I_*)_c = 0.224$ for a mild slope to $(q_* I_*)_c = 0.034(|\Psi(\theta)|)^{5/2}$ for a steep slope. The location of the slope at which the unit discharge is defined and applied to Equation (8) is assumed to be the end of the slope, because the temporal and spatial change of overland flow is supposed to be relatively small, and sediment erosion does not sharply respond to the change of flow depth variation.

To specify the value of constant $A_0$ empirically, we used the data from several studies. Figure 3 presents the relationship between the non-dimensional transport rate $q_{B_0*}$ on bare land and the non-dimensional stream power $(q_* I_*)$, and the results of previous experiments. We used the data of erosion rate in a nearly flat bed. Since soil aggregates were used, the diameters were taken to be the mean. The solid line in Figure 3 expresses the equation with $K_0 = 1.25$, $m = 5/3$, and $(q_* I_*)_c = 0.002$; the sediment erosion rate can be derived reasonably instead of estimating $(q_* I_*)_c$. The discrepancy in Figure 3 comes from the raindrop impact, which has a great influence on structure of rain-impacted flow, which was not included in Equations (2) and (3). Equation (8) indicates a rough trend of sediment transport rate due to rainfall runoff, and suggests that $K_0$ remains constant for sediment erosion by raindrop-impacted flow. Under the assumption of constant rainfall intensity ($r_0$), the kinematic wave runoff model shows that the unit discharge flow $q = r_0 L$, where $L$ is slope length. This means that the stream power $qI = r_0 L I$ is equivalent to the classical index of the product of rainfall intensity, slope length, and slope gradient.
2.3 Sediment Erosion Rate with Particular Reference to Land-Use

Equations (1) and (8) can be applied to a bare catchment slope. However, since an actual slope has a land-use, the total sediment yield should be modified as follows with reference to land-use:

\[
Q_B = \sum_l q_{Bi} \cdot L_i \cdot \sqrt{(\sigma / \rho - 1)gd_i^3}
\]

\[
= \sum_l \varepsilon \cdot \pi \cdot \gamma \cdot A_0 \cdot \{(q_iL_i)_c - (q_iL_i)_e\}^{5/3} \cdot L_i \cdot \sqrt{(\sigma / \rho - 1)gd_i^3}, \tag{9}
\]

where \(Q_B\) = total sediment erosion rate from catchment slope (m³·s⁻¹), subscript \(i\) = land use, \(L_i\) = corresponding reach length, \(\varepsilon\) = projected area of vegetation cover per unit area, \(\gamma\) = reduction rate due to protection works and countermeasures, and \(\pi\) = cropping and management factor.

3. Sediment Pick-Up Model from Floodplain

Sediment erosion from flood plain

The sediments carried from mountains are usually dropped near rivers because of the mild gradients there. Thus, we assume that the sediments do not flow directly into the river, and that previously dropped sediments might be eroded directly by river flow. Further, we assume that the supply from the catchment slopes to the riverside is equal to the riverbank erosion rate.

The sediment erosion rate from riverbanks is simply written as follows:

\[
q_B = A_i d \cdot p_s \cdot \Delta B, \tag{10}
\]

in which \(q_B\) = sediment yield per unit length, \(A_i d = \) sand diameter, \(p_s = \) sediment erosion rate, \(\Delta B = \) width of eroded riverbank, and \(L = \) eroded reach length. Fine sediment diameter \(d = \Delta B,\)
and $L$ are given by a geographic information system (GIS).

Pick-up function of sediment on flood plain

The sediment erosion rate, $p_{s2c}$, of cohesive fine sediment (Sawai, 1977) was modified to give the following:

$$p_{s2c} = p_s \sqrt{\frac{d}{(\sigma / \rho - 1)g}} = F_0 \sqrt{\tau_s \left(1 - \frac{k_p \tau_{sc}}{\tau_s}\right)}^3,$$

in which $F_0$ = empirical constant dependent of water content or soil moisture ($1 \times 10^{-4} - 8 \times 10^{-3}$), $k_p$ = empirical constant ($= 0.4$), and $m = 3$. Figure 4 compares Equation (11) (when $F_0 = 1 \times 10^{-4} - 8 \times 10^{-3}$) and measured data (Sawai, 1977).

Since this study uses the flow intensity of unit stream power, the above expression is modified into the following form:

$$p_s = \sqrt{\frac{(\sigma / \rho - 1)g}{d}} F_0 \left\{ \frac{q l}{\phi(\sigma / \rho - 1)\sqrt{(\sigma / \rho - 1)gd}} \right\}^{4/5} \left\{1 - k_p \frac{(ql)_c}{ql}\right\},$$

in which $\phi$ = velocity coefficient and $(ql)_c$ is estimated by means of Equation (4) for an alluvial riverbed.

![Figure 4](image)

Figure 4  Pick-up rate of cohesive fine sediment.

4. Use of Sediment Erosion Model to Jialingjiang Catchment

4.1 Study Area

In the Jialingjiang catchment, four sub-catchments (I to IV in Figure 2) were delineated with a digital elevation model (DEM) and the digital river network data in GIS, Arc Info 8.2, for sediment runoff modeling. We used a DEM with a spatial resolution of 1 km derived from
the Global Land One-km Base Elevation (GLOBE) Project (GLOBE Task Team, 1999). The digital river network was produced by the Institute of Geographical Science and Natural Resources Research, Chinese Academy of Sciences, from 1:50 000 and 1:100 000 topographic maps covering the catchment. Table 1 presents area, mean elevation, and mean slope for each land cover type in each sub-catchment.

Figure 5 shows the proportions of catchment slopes occupied by farmland. Sub-catchments I and II cover land use in mountainous regions, sub-catchment III covers land use on the plain, and sub-catchment IV covers continuous land use from the mountains to the plain.

Table 1  Topographic properties of sub-catchments I to IV.

<table>
<thead>
<tr>
<th>Landcover</th>
<th>Sub-catchment I (Total area: 4961km²)</th>
<th>Sub-catchment II (Total area: 4063km²)</th>
<th>Sub-catchment III (Total area: 283km²)</th>
<th>Sub-catchment IV (Total area: 3709km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Area (km²)</td>
<td>Mean elevation (m)</td>
<td>Mean slope (-)</td>
<td>Area (km²)</td>
</tr>
<tr>
<td>Forest</td>
<td>2288</td>
<td>3143</td>
<td>0.262</td>
<td>3561</td>
</tr>
<tr>
<td>Bush &amp; shrub</td>
<td>1133</td>
<td>3063</td>
<td>0.240</td>
<td>524</td>
</tr>
<tr>
<td>Paddy field</td>
<td>0</td>
<td>-</td>
<td>-</td>
<td>0</td>
</tr>
<tr>
<td>Farmland</td>
<td>302</td>
<td>2336</td>
<td>0.266</td>
<td>357</td>
</tr>
<tr>
<td>Grassland</td>
<td>4262</td>
<td>3345</td>
<td>0.220</td>
<td>3190</td>
</tr>
<tr>
<td>Urban area</td>
<td>0</td>
<td>-</td>
<td>-</td>
<td>0</td>
</tr>
<tr>
<td>Wasteland</td>
<td>418</td>
<td>3787</td>
<td>0.130</td>
<td>526</td>
</tr>
</tbody>
</table>

Figure 5  Land use of catchment slope according to slope gradient.

Figure 6  Mass movement diagram (modified from Peltier’s original diagram).
Since sediment yield in monsoon Asia is characterized by unsteady and intermittent transport due to intensive mechanical and chemical weathering, it is difficult to use a surface erosion model to estimate sediment yield. Peltier (1950) proposed a diagram that indicates the probability of mass movement as a function of weathering. Weathering processes are represented by annual average temperature and precipitation. The annual average temperature and precipitation of each sub-catchment are plotted in Figure 6. As shown in Figure 6, the reason why sub-catchments I, II, III, and IV were selected is that each has different hydro-geomorphologic characteristic. Sub-catchments I and II experience the least intense erosion, sub-catchment IV experiences the most intense erosion, and sub-catchment III experiences moderate erosion. This makes it possible to apply the surface erosion model successfully to sub-catchments I and II. Applying the model to the four sub-catchments and comparing the results with observations will reveal how well the surface erosion model can be used to model sediment yield in monsoon Asia.

4.2 Input Data

Surface runoff volume must be simulated before the application of the sediment runoff model. We used the same method as that applied to the whole upper region of the Changjiang basin by Hayashi et al. (2003) to simulate the hydrological processes, including surface runoff volume, from each sub-catchment. The hydrologic process of HSPF was applied to the lumped area (HRU) for each land cover type in each sub-catchment. We classified the land cover into seven types: forest, bush and shrub, grassland, farmland, paddy field, urban area, and wasteland. To calculate the volume of surface runoff from each HRU, HSPF needs the slope length of each HRU. The length of one side of an HRU was used as slope length by assuming each HRU to be a square.

Daily precipitation data from 431 observatories regulated by the Changjiang Water Resources Commission in 1987 were used for this study. We used the universal kriging method to convert the observed point data to distributed data with a spatial resolution of 0.5° × 0.5°. A linear model was selected as the theoretical semivariogram for convenience sake. For air temperature, dew point temperature, wind speed, and solar radiation, we used 6-hour data for 1987 with a regular spacing of 1° (lat/long) developed by the ISLSCP (Sellers et al., 1995). Spatially distributed potential evaporation data were produced by Penman's equation from the ISLSCP data. The data were converted from raster to vector format by overlaying each sub-catchment polygon on the 1° grid data and using the area-weighted mean method. The time interval was changed from 6 hours to 1 hour for model calculation by equal division for precipitation and linear interpolation for the other items. To remove the effect of the initial value of storage in the soil layer and volume in each stream reach, the actual calculation was carried out after preliminary calculations for two years by using this meteorological input data two times.

To evaluate the simulated results in each sub-catchment, we used the daily streamflow rate and suspended solids (SS) concentration recorded by the Changjiang Water Resources Commission at the outlet of each sub-catchment (Figure 2).

Calibration of model hydrologic process

The hydrologic process of HSPF was applied to every HRU to select the values of each model parameter so as to satisfy the following criteria in the simulation of runoff volume:

- The absolute value of error in total runoff volume for the calibration period is <10%.
- The absolute value of error in the mean of the low-flow-recession rates based on the computed ratios of daily mean flow today divided by the daily mean flow yesterday for each day is within 0.01 for the highest 30% of the ratios <1.0.
The absolute value of error in the mean of the lowest 50% of the daily mean flows is <20%.

The absolute value of error in the mean of the highest 10% of the daily mean flows is <30%.

The absolute value of seasonal volume error, June to August runoff volume error minus December to February runoff volume error, is <30%.

These error terms were computed by the Hydrological Simulation Program Expert System (HSPEXP). As the above rules of criteria, the default values in HSPEXP were used. For the 10 most sensitive parameters identified (Table 2, 3), we used the initial values approximated from the results of the application of HSPF to the whole upper region and from previous studies (Srinivasan et al., 1998; Al-Abed and Whiteley, 2002). Each parameter value was selected by manual calibration according to the advice offered in HSPEXP (Bicknell et al., 1997).

Table 2  Important parameters for hydrologic component of HSPF.

<table>
<thead>
<tr>
<th>Identification</th>
<th>Description</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>INFILT</td>
<td>Index zone nominal storage</td>
<td>mmih</td>
</tr>
<tr>
<td>IRC</td>
<td>Interflow recession parameter</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>INTFW</td>
<td>Interflow inflow parameter</td>
<td></td>
</tr>
<tr>
<td>UZSN</td>
<td>Upper zone nominal storage</td>
<td>mm</td>
</tr>
<tr>
<td>LZSN</td>
<td>Lower zone nominal storage</td>
<td>mm</td>
</tr>
<tr>
<td>LZEETP</td>
<td>Lower zone ET parameter</td>
<td></td>
</tr>
<tr>
<td>AGWRRC</td>
<td>Basic ground water recession rate</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>KVARY</td>
<td>Ground water recession flow</td>
<td>mm³</td>
</tr>
<tr>
<td>INFEXP</td>
<td>Exponent in the infiltration</td>
<td></td>
</tr>
<tr>
<td>INFIELD</td>
<td>Ratio between the maximum and mean infiltration capacities</td>
<td></td>
</tr>
</tbody>
</table>

* Dimensionless.

The values of this coefficient vary with each month.

Table 3  Used numerical values in the HSPF model.

<table>
<thead>
<tr>
<th>Selected hydrologic calibration parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parameter</td>
</tr>
<tr>
<td>-----------</td>
</tr>
<tr>
<td>INFILT</td>
</tr>
<tr>
<td>IRC</td>
</tr>
<tr>
<td>INTFW</td>
</tr>
<tr>
<td>UZSN</td>
</tr>
<tr>
<td>LZSN</td>
</tr>
<tr>
<td>LZEETP</td>
</tr>
<tr>
<td>AGWRRC</td>
</tr>
<tr>
<td>KVARY</td>
</tr>
<tr>
<td>INFEXP</td>
</tr>
<tr>
<td>INFIELD</td>
</tr>
</tbody>
</table>

* Range of values for each parameter represent the differences in land cover type within each segment.

4.3 Calculated Results

Figure 7 compares the simulated and observed daily runoff depths in each sub-catchment, and supports the applicability of the Stanford Watershed Model.

One way to evaluate sediment yield from catchment slopes is to measure the suspended
load and wash load at hydrological stations in first-order streams in the Jialingjiang catchment. We applied the model to the wash load observed at hydro-stations along first-order streams in 1987 as shown in Figure 2. Table 4 shows the geometric properties of each catchment upstream of first-order streams, the proportions of land use, the effect of the reduction of vegetation cover, and the effect of protection measures, which were used in the calculations of the model. The stream power was estimated as the product of slope gradient and surface flow discharge at the end of each catchment slope.

Figure 7  Simulated and observed daily river flow discharge in each-sub-catchment.

Table 4  Watershed slope and riverbank properties.

<table>
<thead>
<tr>
<th>Basin</th>
<th>A(km²)</th>
<th>H(m)</th>
<th>θ (°)</th>
<th>L(km)</th>
<th>Movement</th>
<th>i₀</th>
<th>B_max(m)</th>
<th>B_min(m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>8403</td>
<td>1540</td>
<td>12.9</td>
<td>167.9</td>
<td>Low</td>
<td>0.0098</td>
<td>50.3</td>
<td>21.4</td>
</tr>
<tr>
<td>II</td>
<td>8148</td>
<td>898</td>
<td>16.2</td>
<td>177.9</td>
<td>Low</td>
<td>0.0096</td>
<td>51.1</td>
<td>35.3</td>
</tr>
<tr>
<td>III</td>
<td>464</td>
<td>17</td>
<td>0.0</td>
<td>30.0</td>
<td>Medium</td>
<td>0.0006</td>
<td>178.0</td>
<td>0.8</td>
</tr>
<tr>
<td>IV</td>
<td>6374</td>
<td>287</td>
<td>8.4</td>
<td>162.4</td>
<td>High</td>
<td>0.0074</td>
<td>143.0</td>
<td>35.5</td>
</tr>
</tbody>
</table>

Since the representative sediment diameter in each catchment was unknown, we assumed the value of d to be 0.0025 m, which has a significant influence on sediment yield.

In calculations of the Kusaka–Fujita model, the runoff ratio f is assumed to be 0.5 for each sub-catchment on the basis field data (Higgit, 2001).

In calculations of the unit stream power model, it was assumed that A₀ = 1.01 for each
catchment when \( K_0 = 10 \), \( \phi = 4 \), and \((q_f I)_c = 0.002\).

In calculations of the sediment pick-up model in the flood plain, we assumed an aggregate soil with low water content, and \( F_0 = 8 \times 10^{-3} \). \((q_f I)_c\) in Equation (11) was determined from observed data. In each sub-catchment, the flow discharge that transported \( Q_b = 10 \) kg/s was taken to be the critical discharge, and converted into the critical unit stream power \((q_f I)_c\). Each \((q_f I)_c\) was converted into dimensionless tractive force: \( \tau_c = 0.17 \) for sub-catchment I, \( \tau_c = 0.145 \) for II, \( \tau_c = 0.0023 \) for III, \( \tau_c = 0.080 \) for IV. The width of floodplain covered by sediment was assumed to be half of the difference between maximum and minimum widths.

![Figure 8](image1.png)

**Figure 8** Comparison between calculated and observed annual sediment yields in each sub-catchment.

Figure 8 compares measured and calculated annual loads for each of the three types of sediment erosion and yield model. This Figure indicates that annual sediment erosion and yield can be estimated by means of surface erosion models whose erosion agents are raindrop impact, overland flow, and river flow, and which explain that the sediment from catchment slopes is transported to near the river and is then flushed out by river flow.

Accordingly, for the purpose of estimating sediment load in a river, it is important to estimate sediment erosion and yield on a catchment slope. In the Kusaka–Fujita model (modified USLE type), the calculated sediment erosion depends on the accuracy of the runoff ratio \( f \), and consequently the calculated value in sub-catchment IV (flat plain) is much greater than the measured value.

When we estimate the sediment load in a dam, we need at least the monthly inflow of wash load into the reservoir, and sometimes the daily inflow.

Figures 9a–d compares the calculated and measured monthly sediment yields. Figure 10 shows the daily surface runoff depth, and the concentrated surface runoff for specified rainfall by comparison with Figure 10. In the unit stream power model, since the surface runoff occurred only during the rainy season, sediment yield occurred only during this season. In the raindrop impact model the sediment yield occurred throughout the year. This indicates that the two models give different types of storage on catchment slopes and transport from slopes to river. Because these two models give equal annual sediment yield, it is better to use the unit stream power model to estimate the sediment erosion during the heavy rainy season. The pick-up model best expressed the monthly trend in the flood plain, especially the increasing trend from spring to the beginning of summer. However, it produced a small discrepancy between the calculated and measured monthly yields, so to find out the cause we compared the calculated and measured daily yields as shown in Figure 11.
Figure 9  Comparison between calculated and observed monthly sediment yields in each sub-catchment.

Figure 10  Simulated and observed daily surface runoff depth in sub-catchment II.

According to the mass movement diagram, we expected the surface erosion model to be applicable to sub-catchments I, II, and III. As shown in Figures 11a–c, the pick-up model in
the floodplain indicates the approximate daily variation up to the beginning of summer. Figure 11d indicates the temporal lag between the calculated and measured daily yields after the summer rainy season. We suppose this to be due to the fact that we assumed that sufficient volume of sediment exists along the floodplain and that all the sediment that the river flow intensity (unit stream power) can transport is picked up. In sub-catchment II, the vast amount of sediment during summer indicated by the unit stream power model cannot reach the river region, and hence not enough sediment is picked up along the floodplain. The sediment yield after the summer rainy season is gradually transported into the river area downpour by downpour, and sufficient sediment is transported up to the next spring, so the model assumption would be valid.

![Figure 11](image)

**Figure 11** Comparison between calculated and observed daily sediment yields in each sub-catchment.

Sub-catchment IV is located near the most intense region in the mass movement diagram. Hence, we suppose that the surface soil layer is saturated and becomes unstable because of heavy rain in summer, and afterwards even relatively weak rain can cause unsteady and intermittent sediment yield such as through surface layer slide, landslide, and land collapse, and so the yielded sediments are transported to near the river. Since the temporal lag between the observed and calculated daily loads is due to the unsteady sediment yield, such a mass movement type of sediment yield goes beyond the assumptions of proposed surface erosion and yield models, which require a unique relationship between external driving forces and sediment erosion and yield.

These calculated results indicate that for the purpose of estimating annual sediment yield in the Jialingjiang catchment, the surface erosion and yield models could be applied irrespective of what the mass movement diagram indicates. For the purpose of estimating
monthly sediment yield, the pick-up model in the floodplain is appropriate. If we use a USLE-type model, we have to increase the precision of the runoff factor, which means the necessity of coupling USLE with a rainfall runoff model.

To improve the precision of estimation of monthly sediment yield and simulated daily sediment yield, we have to study two factors: the mechanism of unsteady and intermittent sediment yield, and the temporal and spatial lag between sediment yield on a catchment slope and sediment erosion and pick-up along the floodplain.

To develop this type of sediment yield model, we need to consider how we can include the probabilistic relationship between local gradient and deposition, the unsteady intensity of averaged rainfall, and representative slope. However, our results support the view that the proposed model can be used to evaluate monthly and annual sediment yields.

5. Conclusion

The control of sediment throughout a catchment is indispensable to managing water environments. For the purpose of predicting sedimentation problems in the upstream Changjiang and in the Three Gorges Dam reservoir, it is important to study which types of erosion model can be used, so we investigated three types of surface erosion and yield model.

To verify candidate models, we applied them to sediment yield in the Jialingjiajiang in 1987, and investigated their applicability and limitations. Although the Jialingjiajiang catchment is so wide that sediment runoff is likely to be much influenced by multiple scales of space and time, the best model could explain the sediment erosion to some extent.

The three models tested approximately simulated the annual sediment yield, on the assumption of spatially and temporally continuous transport from catchment slopes to the river. Although the pick-up model in the flood plain roughly explained the monthly and daily variations in sediment yield, the introduction of a mechanism to explain the unsteady and intermittent sediment yield would improve the precision of estimation below the monthly scale.

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Simulation of Light Environment in Forest

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Abstract
In this report, we measured three-dimensional structure of forest using a laser scanner and simulated light environment of forest. Forest structure was represented by three-dimensional array of cell and each cell was characterized with leaf area index (LAI) and average leaf inclination angle (ALIA). Light environment of forest was simulated by montecarlo raytracing and the distribution of absorption of light, downward light intensity and upward light intensity were calculated. Furthermore, canopy spectral reflectance was calculated.

Keywords: Laser scanner, Japanese larch, Leaf area index, Average leaf inclination angle

1. Introduction

Remote sensing is one of the effective tools to measure fixation of CO₂ by forest which has a large area of terrestrial surface, because it is superior in objectivity, repetition, and wide viewing. In remote sensing, after the incident sunlight is absorbed / scattered by leaves within the canopy, qualitative change of sun light reflected outside the canopy was measured. Therefore, in order to understand remote sensing data, it is necessary to describe the structure (distribution of a leaf) of forest correctly, but, in old research, the structure of forest has been described only uniformity or vertical distribution in space, and it was not described as three-dimensional distribution. However, in the actual forest, it is known that a leaf distribution is not spatially uniform and the photosynthesis activity of a leaf is also not spatially uniform. Therefore, in order to estimate the amount of CO₂ fixation of a forest correctly using remote sensing, the measurement of the three-dimensional structure of forest and clarifying absorption / scattering process of sun beam within a forest is indispensable.

In this report, we describe the three-dimensional structure of forest measured by the laser scanner in the second section, and the simulation of light environment in forest using the three-dimensional structure of forest in the third section.

2. Modeling the Forest Structure

2.1 Materials and Methods

2.1.1 Study Site
In Tomakomai Flux Research Site (hereafter referred to FRS) in the national forest in Tomakomaishi, Hokkaido (42°44'N, 141°31'E), measurement was conducted on Japanese larch (Larix kaempferi Sarg.) in 2003. With regard to Japanese larch in FRS, the age was about 47 year-old, the height was 18m and planting density was 681.4 plants/ha.
2.1.2 Laser Scanner

The Laser scanner was made from a laser range finder and two rotate stages. Because the laser range finder can only measure distance, the laser range finder is mounted on two rotation stages (ARS-936-HP & ARS-136-HP, Chuo-Seiki), which are attached to perpendicular each other, to measure three-dimensional coordinate from azimuth and zenith angle of laser beam. Wavelength of laser beam is 970 nm and divergence angle is 3 mrad.

2.1.3 Theory

We decide that leaf area index (LAI) which is the leaf area per unit surface area, and average leaf inclination angle (ALIA) which is the angle between leaf surface and horizontal plane, as the parameters representing forest structure. It is known that these parameters strongly affect the absorbing and scattering process of sun light in forest. LAI and ALIA were calculated from Gap Fraction by

\[ T(\theta) = \exp\{ -L \cdot G(\theta) / \cos \theta \}. \]  

(1)

Where, \( T(\theta) \) is Gap Fraction at the zenith angle \( \theta \), \( L \) is LAI, \( G(\theta) \) is the mean projection of unit area of leaf onto a plane perpendicular to \( \theta \) and is related to ALIA. A technique to calculate LAI and ALIA from Gap Fraction was described by Miller (1967), Lang (1986), Norman and Campbell (1989), and in this study, we used the method of Norman and Campbell.

2.1.4 Measurement

In order to reconstruct three-dimensional structure of forest, the forest was described three-dimensional array of cells, and LAI and ALIA was defined in each cell. The size of each cell was determined by dividing the 40 m x 40 m observation area set up in FRS into 8 m x 8 m horizontal division, and dividing each horizontal division into a 1m layer. Measurement with a laser scanner was performed on 36(6 x 6) points which is the corner of each horizontal division. With regard to the shooting direction of laser beam, zenith angle ranged from 10° to 70° with interval of 5°, and azimuth angle ranged from 0° to 360° with interval of 0.4°, so 11700 laser beams were shot by one measurement. The measurement of LAI and ALIA was performed mostly one month from April to December in 2003.

2.2 Results

A vertical profile of LAI measured in August is shown in Figure 1. LAI in each layer was averaged by each of 25 horizontal divisions, and error bar means standard deviation. A profile of LAI has two maximum peaks at the height of 10.5 m and 2.5 m, and it is considered that these peaks represent larch and low trees respectively. The average and standard deviation of LAI which were integrated in the height by each of 25 horizontal divisions are shown as a seasonal change from April to December in Table 1. Integrated LAI increased from April to September, and decreased gradually towards December, finally became almost the same LAI of April in December. As correlating with change of LAI, ALIA decreased, i.e. inclination angle of leaves became horizontally, from April to August, and increased towards December.
Figure 1  A vertical profile of LAI measured in August. The error bar means standard deviation.

Table 1  A seasonal change of LAI and ALIA from April to December.

<table>
<thead>
<tr>
<th></th>
<th>LAI Average</th>
<th>S.D.</th>
<th>ALIA Average</th>
<th>S.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Apr.</td>
<td>3.6</td>
<td>0.4</td>
<td>57.4</td>
<td>10.9</td>
</tr>
<tr>
<td>May</td>
<td>4.1</td>
<td>0.4</td>
<td>51.2</td>
<td>8.6</td>
</tr>
<tr>
<td>Jun.</td>
<td>5.2</td>
<td>0.6</td>
<td>41.7</td>
<td>13.0</td>
</tr>
<tr>
<td>Aug.</td>
<td>5.5</td>
<td>1.2</td>
<td>41.1</td>
<td>10.4</td>
</tr>
<tr>
<td>Sep.</td>
<td>5.7</td>
<td>1.0</td>
<td>46.3</td>
<td>8.7</td>
</tr>
<tr>
<td>Oct.</td>
<td>4.5</td>
<td>0.7</td>
<td>50.9</td>
<td>7.7</td>
</tr>
<tr>
<td>Dec.</td>
<td>3.9</td>
<td>0.6</td>
<td>53.8</td>
<td>4.3</td>
</tr>
</tbody>
</table>

3. Simulation of Light Environment

3.1 Materials and Methods

3.1.1 Canopy Model

Canopy model to simulate the absorbing and scattering process of sun light was defined as follows. The distribution of LAI and ALIA as the structural parameter of forest represents three-dimensional array of cells derived from the laser measurement. Absorbing and scattering coefficient of sun light as the optical parameter of forest was measured in August 2003, and the result was shown in Figure 2. We assumed that spectrum reflectance of the leaf in Figure 2 was the same in all cells, and performed the following simulations.
3.1.2 Montecarlo Raytracing

In raytracing, sunlight was expressed as a set of sun beams and the simulation of the light environment in a forest was carried out by calculating whether each beam is intercepted by the cell or it penetrates. Sun beam was expressed in distinction from direct and diffuse beam, and was defined by vector, intensity, and the starting point to each light. About direct beam, the vector was calculated from the latitude and longitude of a site and the time that performed observation. The normal vector of diffuse beam was defined as the normal vector of polygon, which was determined to divide the sky hemisphere into 48 regions, 6 divisions with respect to zenith angle and 8 divisions with respect to azimuth angle. The ratio of direct and diffuse beam was set to 8:2, and the intensity of diffuse beam of each polygon was determined according to zenith angle. The starting point of beam was given at intervals of 10 cm on the upper horizontal plane of a forest model. Sun beams passed the cell, when the random number generated between 0 and 1 was smaller than Gap Fraction calculated from the equation (1), and when it was larger, they judged with being intercepted in a cell. The sun beams that passed the cell were calculated similarly in the next cell. On the other hand, the sun beams that intercepted in the cell were absorbed or scattered according to the optical parameter of leaf in the canopy and the scattered sun beam in the cell became diffuse beam. The end of calculation was defined that the intensity of diffuse beam was less than threshold or the diffuse beam escape to the sky.

3.2 Results

The result of a light scattering simulation is shown in Figure 3. Figure 3(a) shows that the peak of absorption of sun beam was between 11 m and 12 m in height, and was seen in the place higher than the peak of LAI which was shown in Figure 1. Moreover, in spite of having seen the distribution of LAI also in the lower layer, absorption of sun beam was hardly seen in a lower layer. Figure 3 (b) and Figure 3 (c) show the relative intensity of downward and upward sun beam respectively, when the intensity of sun beam at the top of the tree crown is 100%. Both of the Figures show that the intensity of sun beam rapidly decrease from the height to which LAI becomes large, and light has hardly reached a lower layer. Figure 4 shows the spectrum reflectance of forest calculated by simulation, and the spectrum
reflectance of forest is shown for comparison. Although a spectrum curve of forest showed the almost the same change as that of leaf, it was quite small in the absolute value.

Figure 3 (a) Absorption of sun beam, and the relative intensity of (b) downward and (c) upward sun beam by Montecarlo raytracing simulation.

Figure 4 A spectrum reflectance of the forest (solid line) and the leaf (dotted line) by Montecarlo raytracing simulation.

4. Discussion

With regard to seasonal change of LAI and ALIA, LAI increased as the leaf grew thick, and ALIA became small. About seasonal change of ALIA, since it is influenced by stem and branch at a fallen-leaves term, the average leaf inclination angle gradually approaches perpendicular, but in order to be influenced by a horizontal leaf as a leaf grows thick, the average leaf inclination angle approaches horizontally. The result of a simulation showed that sun beams reach understory by about 3% of the top of the canopy. The simulation has calculated 10 m x 10 m averaged sun beam, and is not taking the influence of sun fleck into
consideration. Photosynthesis of lower layer vegetation is told that sun fleck has a big contribution, and it is necessary to evaluate the influence of sun fleck in an actual forest, from now on. As compared with the spectrum reflectance of leaf, it turns out that the spectrum reflectance of canopy is quite low. Because a lot of the incident light was absorbed by multiple scattering in the canopy compared with the leaf. In future, in order to estimate the amount of photosynthesis of a forest, it is necessary to include the photosynthetic response to light environment within a canopy in the simulation.

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Publications and Presentations

Original Papers and Reviews:

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Overview of the NIES
Supercomputer Systems
Overview of the NIES Supercomputer Systems

Akihiro MUSA, Masahiro MIYAZAWA, Nanaumi NAGAMINE and Yoshikazu SATOH
NEC Corporation, HPC System Center

1. System

1) Introduction

In February 2002 National Institute for Environmental Studies (NIES) installed the NEC SX-6 systems that substitutes for the SX-4/32, the SX-6/64M8 with the operating system SUPER-UX. Figure 1 shows the system configurations with the SX-6 (for Vector calculate server) and the other central machines; NEC Express5800/1160Xa (for Scalar calculate server and Front-end server), Compaq GS160, SGI Octane2, SUN Blade 1000, and so on. The network is based on Gigabit Ethernet Switch.

The SX-6/64M8 provides a peak vector performance of 512GFLOPS (64GFLOPS per node), and has the following features:

- 512Gbytes of main memory (64GB per node)
- 8TBytes of raid disk capacity with Global File System (GFS)
- 200TBytes of Tape Library capacity (DTF-2 format)
- Internode crossbar switches (IXS) interface at 8Gbps
- 1000Base-SX interface at 1Gbps

The SX-6 and the Express5800/1160Xa use the GFS, which provides high-speed internode file sharing.

Figure 1 System configurations.
2) Feature of the SX-6 System

Based on experience with the SX-4 Series, the SX-6 Series supercomputers have been developed as a system that aims at considerably improved cost-performance and covers the high-end computing range at a level above the conventional machines, while minimizing software development and operation costs and pursuing both ease-of-use and high effective performance.

Inheritance and expansion of the distributed shared memory architecture

The SX-6 Series inherits the vector-processor (CPU) based distributed shared memory architecture which was highly praised in the SX-4 Series, and flexibly works with all kinds of parallel processing schemes. Each shared memory type single-node system contains up to 8 CPUs, which share a large main memory of up to 64Gbytes. In a multi-nodes system, configured with a maximum of 128 nodes, parallel processing by 1024 CPUs achieves vector performance of 8TFLOPS and provides a large-capacity memory of 8Tbytes, making it possible to flexibly handle large-scale computing requirements.

Through inheritance of the SX architecture, the operating system (SUPER-UX: 64-bit UNIX enhanced for supercomputers) maintains perfect compatibility with the SX-4 Series, and the highly rated flexible operating configurations can be used as is. In addition, software development and operating environments (PSUITE, SPINE ware) can also be used without any changes.

Full-featured application software

Many kinds of application software finely tuned for the SX Series can be used on the SX-6 Series without modification. This means that high effective performance in a wide range of fields can be achieved without the inconvenience of further tuning.

Operation cost performance considerably improved through realization of single-chip vector processor

Used in recent SX Series, Complementary Metal-Oxide Semiconductor (CMOS) technology enabled development of the single-chip that can be higher performance than the vector processor (scalar unit and vector unit) consisting of dozens of LSIs.

By using the developed single-chip, the SX-4 Series processor has improved cost performance such as power consumption, heat generation and floor space.

As compared with the SX-4 Series processor with a maximum performance of 2GFLOPS, the SX-6 Series processor achieves operability performance 4 times by speeding up the machine cycle and Using single-chip processor.

High scalability

The SX-6 Series, by realizing scalable balance between operation using 8 CPUs in the node and memory performance, achieves high effective multiprocessing performance and scalability. The Series flexibly responds to a wide range of needs with configurations of up to 1024 CPUs, by connecting the nodes using 2 to 128 internodes crossbar switches (IXS).
2. Operation

1) Management situation
   The job class was changed as follows in consideration of the setup of daily management situations or the priority in the FY2003.

   - present

<table>
<thead>
<tr>
<th>Job class</th>
<th>CPU limit</th>
<th>Memory limit</th>
<th>Run limit</th>
<th>Execute nodes</th>
</tr>
</thead>
<tbody>
<tr>
<td>vector_sd</td>
<td>0.5H</td>
<td>16GB</td>
<td>2</td>
<td>1nodes</td>
</tr>
<tr>
<td>vector_s3h</td>
<td>3H</td>
<td>16GB</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>vector_s24h</td>
<td>24H</td>
<td>16GB</td>
<td>5</td>
<td>1nodes</td>
</tr>
<tr>
<td>vector_s15d</td>
<td>360H</td>
<td>16GB</td>
<td>11</td>
<td>2nodes</td>
</tr>
<tr>
<td>vector_ps</td>
<td>192H</td>
<td>60GB</td>
<td>1</td>
<td>4nodes</td>
</tr>
<tr>
<td>vector_multi (vector_atm)</td>
<td>5760H</td>
<td>60GB</td>
<td>1</td>
<td>2nodes</td>
</tr>
</tbody>
</table>

2) Improvement of system use
   The following policies have made it possible to achieve efficient use of the system.

i) Plentiful RAS Functions
   In the SX-6 Series, a dramatic improvement in hardware reliability is realized by using the latest technology and a high-integration designs such as the single-chip vector processor, while further reducing the number of parts. As conventional machines, error-correcting codes in main memory and error detecting functions such as circuit duplication and parity checks have been implemented. When a hardware error does occur, a built-in diagnosis function (BID) quickly and automatically indicates the location of the fault, and an automatic reconfiguration function releases the faulty component and continues system operation. In addition to the functions above, prompt fault diagnosis and simplified preventive maintenance procedures —using automatic collection of fault information, automatic reporting to the service center, and remote maintenance from the service center— result in a comprehensive improvement in the system’s reliability, availability and serviceability. Such measures have improved the efficiency of the multi-nodes system.
Association (sharing files) with the IA-64/Linux server using the GFS (Global File System), which provides high-speed internode file sharing (Figure 2).

![Diagram](image)

**Figure 2  Association with IA-64/Linux Server through GFS**

**ii) Support of program improvement in the speed**

The example of the program improvement in the speed performed in the past is shown below.

<table>
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<tr>
<th>Model Name</th>
<th>The contents of correction</th>
<th>Improvement ratio</th>
</tr>
</thead>
<tbody>
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<td>agcm5.5p</td>
<td>Tuning to the routine out of which the performance has not come</td>
<td>1.3</td>
</tr>
<tr>
<td>RAMS Ver4.3</td>
<td>Sauce correction Optimization by the compile option</td>
<td>4.0</td>
</tr>
<tr>
<td>CCSR/NIES AGCM</td>
<td>Parallelizing by sauce correction</td>
<td>5.0</td>
</tr>
<tr>
<td>agcm5.4.02</td>
<td>Parallelizing by sauce correction</td>
<td>2.0</td>
</tr>
<tr>
<td>flood2D</td>
<td>Vector tuning Parallelizing tuning</td>
<td>49.0</td>
</tr>
<tr>
<td>FldDynaNR_Fortran</td>
<td>Use of mathematicallibrary: ASL Sauce correction</td>
<td>40.0</td>
</tr>
<tr>
<td>MJ98-CTM</td>
<td>Parallelizing tuning</td>
<td>2.0</td>
</tr>
</tbody>
</table>

**3) Future plan**

We plan to improve the efficiency by better management and fine tuning program. Information on the system use and operation will be disseminated by a portal site.
3. Use of SX-6

Figure 3 shows supercomputer’s operation results from Apr. 2004 to Dec. 2004.

![Graph showing job count and CPU time from April to December 2004.](image)

**Figure 3** Supercomputers’ operation results.

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## Research Programs in FY2003

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<td>13-20</td>
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<td>A climate simulation at the early stage of the last glacial period (Meteorological Research Institute)</td>
<td>21-30</td>
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<tr>
<td>Development of atmospheric general circulation model for terrestrial planets and related fundamental experiments on the atmospheric structures (Hokkaido University)</td>
<td>77-86</td>
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<td>Atmospheric motion and air quality in East Asia (Kyoto University, NIES)</td>
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Appendices

Outlines of the Research Activity Reports
(in Japanese)

Program of the 12th Supercomputer Workshop
Tsukuba, October 4, 2004
Outlines of the Research Activity Reports
(in Japanese)
亜熱帯西太平洋上のオゾン全量極小の年々変動

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1. 背景と目的

台湾の東に位置する亜熱帯西太平洋域には、冬季に非常に低いオゾン全量の値（オゾンホールの
定義値 220 DU 以下）がしばしば出現する。2001 年の 12 月の値はかなり低く、那覇や香港では最
低値は 190 DU まで減少した（Kawahira, 2002）。そこで、この非常に低いオゾン全量が、季節変動
として毎年起こっているものなのか、あるいは、最近年々この領域のオゾン量が少なくなっている
のか、さらに、上部対流圏～下部成層圏付近の低温域で生じる可能性のある氷などの粒子上で
起こる不均一反応の影響はあるのか、を明らかにするため、EP_TOMS によって観測されたオゾ
ン全量の時系列解析と、CCSR/NIES 化学輸送モデルによる数値実験を行った。

2. EP_TOMS オゾン全量解析とナッシング化学輸送モデルによる数値実
験の結果

1996年から2002年までのEP_TOMSによるオゾン全量の観測データを使い、亜熱帯西太平洋域
（15-25°N、120-150°E）での観測値を面積の重みをつけて平均し、この領域平均に関してZiemkeの
方法（Ziemke, 1997）を使って時系列解析を行った。この時系列解析では、季節変動、太陽活動の11
年周期、QBO、El Nino、対流圏上部～成層圏下部気温（対流圏界面高度などに応答）の影響を調べ
ることができる。その結果、振幅の大きい季節変動成分を除いた残りの年々変動成分について、
この期間中オゾン全量の特に少なかった96-97、98-99、2001-02の冬季は、QBOの成分が大きいこ
とかわかった（図1）。また、これらの年にはこの領域がQBOの特定の位相（約30 hPa以上の高度で
西風、それ以下の高度では東風のいわゆる西風シーア、図2上段の図および下段の図の実線）とな
っており、その場合に駆動される大気循環の上昇域に入るために、オゾン全量が少なくないてい
ることがわかった。

図1 季節変動成分を取り除いたオゾンアノマリの時系列。1996年8月から2002年7月までの解析結果。15-25°N、
120-150°Eの範囲の面積で重みを付けたオゾン全量の平均値に対して解析を行った。オゾン全量アノマリ（黒）、
QBO成分のオゾン全量アノマリへの寄与（緑）、ENSO成分の寄与（水色）、MSU温度の寄与（赤）で示す

—167—
次に、ECMWF気象データを用い、不均一反応過程を除外したCCSR/NIESナッジング化学輸送モデルによる計算を行った。計算では、45°N-45°Sの緯度帯で不均一反応過程を除外した。このモデルが、亜熱帯西太平洋域の12月～1月のオゾン全量の最低値の年々変動をよく再現していることがわかる。また、氫粒子と硝酸3水和物（NAT）による不均一反応過程を導入した数値実験結果と、前述の45°N-45°Sの緯度帯でこれらの不均一反応過程を除外した数値実験結果との差から、1996年以降のこの領域の低いオゾン全量に関しては、不均一反応によるオゾン破壊の影響は多目に観積もっても2～3 DU程度であることがわかった。さらに詳しく解析の結果、この不均一反応の影響の大きさは、赤道縦2年振動（OBO）の影響（5～10 DU）、エルニーニョの影響（5 DU以下）、太陽活動の影響（5 DU）に比べると小さいが、無視できるほど小さい値ではないこと、などがわかった（Zhou et al., 2003）。

図 2（上）27°N 106°EにおけるECMWFデータの東西風速。正の値（実線）が西風、負の値（点線）が東風を表す。等価線の間隔は10 m s⁻¹。 （下）亜熱帯西太平洋域（15-25°N, 120-150°E）のオゾン全量の年々変動。1996年8月から2002年7月までの、面積で重みをつけたこの領域での平均値を示す。実線：EP_TOMSによる観測値。波線：化学輸送モデルによる計算結果。赤い縦波線は、3つのオゾン極小イベントを表す。

3. 結わりに

以上の、EP_TOMSのオゾン全量データの時系列解析と、ナッジング化学輸送モデルを用いた数値実験により、1996-2002年の間の亜熱帯西太平洋域で観測されたオゾン全量の極端な減少は、主にOBOによって誘起された子午面循環によるものであることがわかった。オゾンは低緯度域の成長雲で活発に生成され、この時期のオゾン光化学反応の季節変動（冬期のオゾンの光化学生成が減少）をOBOの影響によって、この時期に極端なオゾン減少が起こったものと考えられ、また化学輸送モデルを用いた数値実験により、NATや水粒子が関与する不均一反応によるオゾン破壊のこの時期の影響は、OBOの影響に比べると少ないと結論されたが、全く無視できるほど小さいわけではない。この領域の下部成層圏の反応性塩素（Cluppy）濃度などの観測が必要である。
CCSR/NIES AGCMにネスティングしたNIES-RAMSによる
21世紀のアジアの水循環変動

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1. 背景と目的
近年、地球温暖化によって水循環はどのような影響を受けるのか、ということに対して人々の多大な関心が集まっており、人間生活、自然生態系にしばしば多大な影響を及ぼす、渇水・洪水・土砂災害・土壌浸食などに対して適切に水資源を管理するために、政策立案者から将来のリスクアセスメントがますます求められるようになっている。
現在、地球環境変動下の水循環予測は、主に全球気候モデルを用いて行われているけれども、現段階では地域スケールの水資源、農業生産活動などの影響評価のために用いることができる情報は非常に限られている。全球気候モデルによってグローバルに予測された気候変動の結果を、地域スケールの気候に適用して水循環の動態を明らかにし、水資源管理に役立てるには、ダウンスケーリングすることが必要となる。
本研究は、地域気候モデルの開発を行い、全球気候モデルの結果を物理法則に基づいてダウンスケーリングすることによって、水平解像度60 kmでモンスーンアジア地域気候に対する地球環境変動の影響の研究を行った。

2. 概要
国立研究所では、地域スケールの水循環変動予測・影響評価のために、Regional Atmospheric Modeling System (RAMS) をベースに、親モデルであるCCSR/NIES AGCMと矛盾のない共通の物理過程を持つ地域気候モデル（NIES-RAMS）を開発してきた。例えば、陸面過程はGCMと同じMATSIROを使っている。本研究では、CCSR/NIES AGCMによる現在気候・将来気候のタイムスライス実験結果をNIES-RAMSの初期・境界条件として与え、それぞれ10年間ずつ積分を行い、地球温暖化時におけるアジアの水循環変動を検討した。

図1 全球気候モデル(CCSR/NIES AGCM)と地域気候モデル(NIES-RAMS)によるモンスーンアジア地域での水平解像度化
ヨーロッパ中期予報センターの再解析値（ERA15）を境界条件とした、NIES-RAMS による現在気候再現実験結果は、再解析値に見られるモンスーン循環、亜熱帯気圧、偏西風の特徴をよく再現できており、高度度も比較的よく再現できていた。ただし、ERA15 に比べ、NIES-RAMS では、高度度が全体に高くなる正のバイアスが見られた。降水分布については、多くの地域気候モデルで見られるように、計算領域の側方境界付近において非現実的で降水が生じていたが、フィリピンから日本に伸びる降水帯や、西太平洋やベンガル湾の強い降水帯などの降水分布を比較的よく捉えることができた。また、解像度の細い CPC Merged Analysis of Precipitation (CMAP) や、Global Precipitation Climatology Project (GPCP) では見られないが、TRMM で見られるインド西岸やチベットの南斜面における地形性降水もモデルで比較的よく捉えることに成功していた。

降水の観測値はデータセットによるばらつきもあるけれども、総じてモデルは熱帯海洋上で降水を過大評価する気味であった。また、中国の東部においても比較的強い降水が生じていた。これは、NIES-RAMS のバイアスであり、他の変数についても検討した結果、現段階の NIES-RAMS には、高優れデータの中等の低い気圧（4-8 hPa）によって収束が生じ、蒸発蒸発気体が増加し、降水過度をもたらしている。多くの降水は大気中で潜熱を解放することによって蒸発を上昇させ、その結果大気の層厚が厚くなり、高優れバイアスが生じていた。

CCSR/NIES AGCM による現在気候・将来気候のタイムスライス実験結果を NIES-RAMS の初期・境界条件として与えた条件では、親モデルと NIES-RAMS 間で結果に大きな不整合が生じることがなく、CMA では不明瞭であったインド西岸やチベットの南斜面における地形性降雨、日本付近の梅雨前線/台風による降水帯が表現され、力学的ダウンスケーリングは概ね成功していた。

また、温暖化した結果、本研究で用いた CMA では海面水温が El Niño に似た状態を示し、活発な対流が東へシフトし、ウォーター循環が変調し、西太平洋域で下降流が強まり、降水の減少が予測される。さらに、モンスーン循環が強まっているとは必ずしも言えなくなけれども、温暖化による水蒸気量の増加によって水蒸気フラックスは増加し、インド西岸、チベット南側の地形性降水の強まりが見られる。西太平洋域における高気圧性の循環場が強まり、フィリピン諸島から東シナ海にかけて水蒸気フラックスの増加が見られ、日本周辺で降水量が増加することが予測される。将来気候において、人為起源の温室効果気体の増加によって生成する対流エネルギーの増加と、気温の上昇に伴う蒸発量の増加、大気中水蒸気の急激な増加（クラウジウス＝クラベイロンの法則）によって、降水が増え、水循環が活発化し、アジアの多くの地域において年平均流出量が増加することが予測される。

3. まとめ

全球気候モデルの物理過程を地域気候モデルに導入し、全球気候モデルと物理過程が整合的な新たな地域気候モデルの開発を行い、力学的ダウンスケーリングを行い、モンスーンアジア地域の水循環の将来予測を行った。地域気候モデルを用いることにより、CMA では明確に表現できない地形性降雨などのメソスケール現象を表現することができた。人為起源の温室効果気体の増加によって温暖化が予測される将来気候において、対流エネルギーの増加と、気温の上昇に伴う蒸発の増加、大気中水蒸気の急激な増加によって、降水が増え、水循環が活発化し、アジアの多くの地域において流出量が増加することが予測される。また、地域によって洪水リスクの高まりや、農業生産への影響が懸念される。

謝辞：本研究は文部科学省科学技術振興調整費プロジェクト「21世紀のアジアの水資源変動予測」（代表：鬼頭昭雄 気象研究所会長）、環境省地球環境研究総合推進費「地球温暖化の総合解析を目指した気候モデルと影響·対策評価モデルの統合に関する研究(IR-3)」（代表：神沢 博 名古屋大学教授）の一部として行われました。
最終氷期初期の気候系シミュレーション

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1. はじめに
最終間氷期(Eemian)から最終氷期への移行は、11万8千年前頃から10万6千年前頃の間に起こったと考えられている。この間に海面水位は約50 m～80 m程度低下し、等量の淡水が氷床の前で陸上に蓄積されたことを示唆している。実際、11万5千年前頃にはカナダのBaffin島やQueen Elizabeth諸島において氷河の形成された痕跡がある(例えばClark et al., 1993)。11万5千年前頃は、北半球夏に地球全体が受ける太陽放射量が過去数万年の間では最小になることが地球軌道要素の計算からわかるが、これは北半球高緯度に涼しい夏と比較的暖かい冬をもたらし、氷床の形成には好条件となることが知られている。しかしながら、大気大循環モデルによる過去の研究では、地表面条件や海面水温を現在の値に固定したまま地球軌道要素のみを変化させても新たな大陆氷床の形成はみられていない。このことは、日射量の変化に伴う地表面・海洋の緩やかな変化が間氷期から氷期への移行に関しては無視できないことを示唆している。ここでは、大気と海洋の相互作用を陽に表現できる大気海洋結合モデルを用いた最終氷期の始まりに関する実験とその経過について報告する。

2. 実験
用いたモデルは気象研究所大気海洋結合モデル(MRIGCM2.3.2)で、境界条件として11万5千年前の地球軌道要素に基づく日射の季節変化(図 1)と産業革命前の二酸化炭素濃度280 ppmvを用いている。また、比較実験として、280 ppmvの二酸化炭素濃度を用いた基準ランを(現在気候値を初期値として)210年間行い、その101年目から115 ka軌道要素による実験を開始した。積分時間は軌道要素の切替以降400年程度であり、気候値としてここではその301年目から400年目までの平均値を、また基準ランでは111年から210年までの平均値を用いた。

![INSOLATION 115ka-0ka,115ka](image)

![ANNUAL MEAN](image)

図 1 115 kaの日射量の季節変化(現在との差)
3. 結果

図2には、115 kaラーンと基準ラーンにおける7月の月平均気温の差を示した。図1からわかるように、115 kaにおける日射は北半球高緯度で少なくなるため、高緯度側の気温は下がるが、低緯度においてはむしろ現在より暖かくなっている。これは115 kaでは冬期から春期にかけての低緯度の日射量が現在よりも増えるため、暖められた海洋の影響が残っているのだと考えられる。つまり図2の温度差は日射量の違いを相関に反映したものとなっている。こうした夏期の気温変化に対応して、積分開始後150年前後からカナダ多島海あたりで越年性の積雪がみられるようになった（図3参照）。これらの地方では夏季の降水が雨から完全に雪に変化しており、年間の正味の降水量は正となっていることが確認された。すなわちこれらの地方では積雪が年々真に増加している。またこの変化が逆に同地点の夏の気温を更に引き下げる効果も確認された。こうした結果は水期初期の氷床形成過程に関する標準的な見解を支持するものであるといえる。
GCMにおける雲の表現の違いが
モデル気候感度に及ぼす影響について

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1. 背景と目的

気候感度（大気中のCO₂倍増に対する地表面気温の平衡応答）は、温暖化シナリオ実験における気候モデルの気温上昇幅を理解する上で重要な指標である。温暖化シナリオ実験の気温上昇幅は異なる気候モデルの間で、同様の放射強制力を与えた場合でも21世紀後半に3℃もばらつくことが指摘されている（IPCC第三次報告書）。モデル間の気温上昇幅の違いは主に気候感度の違いと海洋への熱吸収速度の違いによるため、温暖化予測に対する信頼性を今後高めるためにはモデル間で気候感度がばらつく原因を特定することが望ましい。

本研究では雲過程に注目し、雲のモデル表現がCCSR/NIES/FRCGC AGCMの気候感度にどのように影響するかを議論する。具体的には、AGCMの大規模凝結過程を不確定性の範囲内で変数することにより気候感度4.0℃のモデル（以後、「低感度版」と「高感度版」）を作成した。2つのパーソンの設定の違いがどのような仕組みで気候感度に違いをもたらすかを議論し、気候感度を緩和するプロセスを記述する。温暖化見通しの不確定性を扱えるため今後とも気候モデル改良が必要となるが、その方針を策定する際、以下で得られた結果が参考となるならば幸いである。

2. モデル・数値実験設定

本研究で紹介する数値実験では、CCSR/NIES/FRCGC AGCMを混合層（スラブ）海洋モデルと結合して使用した。「低感度版」と「高感度版」のモデルは雲過程の表現で設定が異なる。即ち、

1. 雲水量のうち液相の占める割合を管理する経験的関数の定義が異なる（図1）。

2. 重力波に伴い崩壊した雲氷は「低感度版」では水粒子として地表へ落下させる一方、「高感度版」では雲氷（水相）として大気中に留める。

「低感度版」と「高感度版」のモデルをそれぞれ標準実験（産業革命前の条件）として45年、CO₂濃度倍増実験として45年積分した。CO₂倍増に対する平衡応答を「低感度版」と「高感度版」で比較するために、積分の最後5年間を準平衡状態と見なしして平均じ解析に用いた。

3. 成果

CO₂倍増に対する正味雲放射強制力の応答を「低感度版」と「高感度版」で比較したところ、低緯度域に見られる放射強制力の減少（温暖化を促進する方向に働く）が「低感度版」では「高感度版」より狭い領域に限定されることが確認できた（図2a-b）。特に南半球中緯度（50-60S）では正味（及び短波）放射強制力の顕著な減少が「高感度版」に見られる（図2b）一方、同様の特徴は「低感度版」には見られない（図2a）。このため南半球中緯度では雲放射強制力のCO₂倍増に対する応答が「低感度版」と「高感度版」で10W/m²も異なり（図2c）、「低感度版」の雲の応答が「高感度版」と比べて温暖化を抑制することが示唆された。

以上で述べた雲放射強制力の応答の違い（「低感度版」vs「高感度版」）は、雲水量の応答の違い（図2d-f）とも矛盾しない。

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4. まとめ

本研究で得られた結果から、モデル気候モデルは雲水/雲水蒸気の生成消滅過程の取り扱いに敏感であることが示された。CCSR/NIES/FRCGC AGCMの場合、
(1) 雲水の相（液相/固相）を診断する経験的関数及び
(2) 重力波に伴う水蒸気の取り扱いを不確定性の範囲内で変更することにより、気候モデルは4.0℃から6.3℃に変化した。経験的関数の変更は、CO₂増加に対する雲水蒸気の緯度/緯度分布を制御し、雲放射制限力の応答を変え、気候モデルに影響を及ぼす。上記プロセスの中で、特に南大西洋上空の雲水蒸気の応答がCCSR/NIES/FRCGCモデルの気候モデルを決定する上で重要な役割を果たすことが示された。

図 1 雲水蒸気（液相＋固相）の中で液相の占める質量の割合を気温から診断する経験的関数。"LOW version"が「低感度版」、「HIGH version」が「高感度版」の設定を相当する。様々なAGCMに採用された関数を重ねて示す。

図 2 CO₂倍増に対する雲放射強制力の応答(a-c, 年・緯度平均)と雲水蒸気の応答(d-f, 年・緯度平均)。a-cの赤線は短波放射強制力、青線は長波放射強制力を示し、黒線は正規（短波+長波）放射強制力を示す。(a), (d)が「低感度版」の応答、(b), (e)が「高感度版」の応答を示す。(e), (f)はCO₂倍増に対する応答の差（「低感度版」－「高感度版」）に相当する。雲放射強制力は上向きを正と定義する。(d), (e)には等温線を標準実験（0℃、-5℃）とCO₂倍増実験（15℃、25℃）について重ねて表示した。
全球雲解像実験を目標とした雲解像モデルによる放射対流平衡

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1. 研究目的
気候モデリングの最大の不確定要素として積雲パラメタリゼーションの問題がある。積雲の運動を直接計算する雲解像モデルを用いれば、これらの困難は回避されると考えられている。最近、全球雲解像モデルが開発された (Tomita and Satoh, 2004)；このモデルは正20面体格子を用いた非静力学モデルである。今のこと全球雲解像実験はごく限られた超並列コンピュータでしか実施できないが、新しいモデルの気候値の信頼性を調べるためには多くの実験を行うなければならぬ。他のモデルやさまざまなコンピュータ上での異なる実験設定との結果などの比較が必要である。

雲解像モデルでどのような気候値が得られるかを調べるために、放射対流平衡実験をさまざまな状況で実施した。熱帯一様条件での実験を行うために、全球雲解像モデルの感帯半径を小さくし、一般的な海面水温をあたえる実験を行った。また、赤道一帯に相当する40,000 kmと緯度方向に100 kmの細長い矩形領域での大規模実験を行い、積雲の組織化について調べた。この報告では、パラメータ依存性を調べるために、より狭い領域での実験を行った。以下では、2重層温度境界条件のもとでのカーテシアン座標系の非静力学モデルの結果を示す (Satoh, 2003)。領域のサイズや格子間隔に対する依存性を調べる。

2. 実験結果
図1に異なる領域、異なる温度解像ののもとで行った実験の鉛直速度の統計値の解析例を示す。これらの図は鉛直速度の最大値の確率密度関数 (pdf) を示している。各時ステップごと鉛直速度の最大値を用いて確率密度関数を計算した。コントロール実験1 (100 km × 100 km, Δx=2 km)の結果では、最大鉛直速度が15-20 m/sとなる頻度を、実験2 (200 km × 200 km, Δx=2 km)よりも小さい。最大鉛直速度の時間平均値は実験1では3 m/sであるのに対して、実験2では10 m/sである。実験2では、pdfの拡がりがより大きくなっており、最大鉛直速度が約20 m/sにまで達している。このこととは、領域のサイズが大きければ、強い上昇流を伴う積雲が領域内に共存し、領域内に強い鉛直速度が頻繁に現れることを表していると考えられる。したがって、pdfは強い鉛直速度の方にピークをもつことになる。さらに領域幅、格子間隔を変えた実験を行ったところ、格子数が100×100を超えると、pdfの形が実験2と似ていることがわかった。この結果は、Tompkins and Craig (1998)で主張されたことと異なっている。つまり、彼らは十分な上昇速度を得るための見積もりか、100 km × 100 kmの領域で格子間隔Δx=2 kmが必要であったのに対し、本実験のpdfの比較によれば、より多くの格子点数が必要であるという結果になった。

図1 放射対流平衡実験での鉛直速度の統計解析。領域内の最大鉛直速度の確率密度関数、1) 100 km × 100 km with Δx = 2 km, 2) 200 km × 200 km with Δx = 4 km.

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気象研究所成層圏化学輸送モデル（MJ98-CTM）による
オゾン QBO について

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1. 背景と目的

熱帯成層圏で約 2 年毎に東風と西風が入れ替わる現象は準二年振動（QBO）と呼ばれ、その影響は熱帯成層圏に限らず高緯度成層圏にまで及んでおり、成層圏の循環にとって非常に重要な要因になっている。QBO に伴い、熱帯での上昇や下降を伴う子午面循環が形成されるので、種々の化学種も大気場の QBO 同じように振動しており、中でも、オゾン QBO は有名である。成層圏の大気場や化学種場の年々変動は QBO に大きく依存しており、長期ランを加えて種々のフォーシングの影響を評価する際にはモデルで QBO を再現することが求められていた。しかし、大気循環モデル（GCM）で QBO を再現することは簡単でなく、ようやくここ 10 年ほどで、いくつか GCM で種々の工夫を施して QBO が再現されるようになった。オゾン QBO の再現はさらに希である。

オゾンや他の化学種の QBO の生成・維持メカニズムは、大気場の QBO に比べ未だ不明の部分が多いけれども、今後のオゾンや他の化学種の変動を定量的に予測していく上で、これら点を解明していくことが不可欠であり、観測からのアプローチと併せて GCM での化学種 QBO の研究が求められている。本研究では、最新の気象研究所の 3 次元化学輸送モデルを用い、化学種 QBO 構造を調べることを目的としている。

2. 概要

気象研究所成層圏化学輸送モデル（MJ98-CTM）は、化学モジュールである大気大循環モデル（MJ98）と化学モジュール（輸送過程と化学過程）をカップリングさせたモデルである。輸送過程は鉛直にはフラックス・フォームのセミ・ラグランジャン法であり、水平には 3 次関数で内挿を行う“普通の”セミ・ラグランジャン法である。化学過程はファミリー法を使い、成層圏の主な化学種を含み、7 種のファミリーを含む 34 の長寿命種、15 の短寿命種、79 の気相反応、34 の光化学反応を扱っている。タイプ I, II の 2 種類の成層成層雲（PSC）と硫酸エーロゾールも含み、PSC 上で 6 種、硫酸エーロゾール上で 3 種の不均一反応を扱っている。

化学モジュールではオゾンを予報しており、その結果は化学モジュールの放射には反映している。つまり、オゾンは Interactive になっている。Baldwin et al. (2001) よると、GCM で QBO を再現するにはおおよそ 3 つの条件が必要であり、それらは対流調節的な対流スキーム、細かい鉛直分解能、小さな水平拡散係数である。気象研究所でもこの条件に沿って GCM を設定し直して QBO を再現した。モデルの水平解像度は T42（経経度約 2.8 度、〜300 km）、鉛直解像度は 68 層（surface〜0.01 hPa）の T42L68 を使った。

対流スキームはデフォルトの 45 層モデル（L45）で使われている従来の

図 1 赤道（5°S〜5°N 平均）上空での帯状平均東西風（生データ）の変化（5〜100 hPa、15 年間）等価線間隔は 5 m/s で、西風は実線、東風は破線。
prognostic Arakawa のままで変更は一切加えていない。代わりに、Hines (1997) のドップラー・スプレッド重力波抵抗スクリムを使い、そのソースを南緯 30 度から北緯 30 度の間でガウス型(半値幅 15 度、振幅 1.0 m/s)を経緯度に独立的に均一ソース(1.5 m/s)に上乗せた。また、対流圈の大循環を変えず、その影響を出来るだけ小さくするという方針で鉛直分解能と水平拡散を L45 の設定から変化させた。

鉛直 68 層の取り方は、150 hPa より上から変え、100 hPa から 1 hPa の間で層厚 500 m、それから上 0.05 hPa の上端まで徐々に増加させた。水平拡散の時定数は高度 150 hPa において最大波数 42 で 18 時間から徐々に増加させ、100 hPa より上で 10 倍の 180 時間とした。

3. 成果

図 1 に 5-100 hPa の带状平均東西風の 10 年分を示す。東西風は約 30 ヶ月で振動しており、西風のシナジーが強く、現実的な OBO を再現している。ただし、観測に比べ、振幅は小さく、特に、西風の振幅は半分くらいになっている。下層ほどその悪評価の程度は大きい。この OBO を作っている加速はモデルで表現される波 (rwv) と重力波 (gwd) の 2 つから来るものである。30 hPa の高さで観ると、rwv 加速は風の変化 9-12 ヶ月先行して正のピークを持ち、このとき gwd 加速が同じ符号だが非常に小さい。gwd 加速はそれから増加し、正のピークになるのは 1-4 ヶ月前で、rwv 加速は殆どゼロである。言い換えれば、rwv 加速が gwd 加速に約 8 ヶ月先行している（図略）。

オゾンのアノマリーを描くと図 2 のように、15 hPa を境にして符号が異なっており、観測値に似ている。15 hPa より下では西風シナジーが強いところで濃いオゾンが出現し、東風の中心で薄いオゾンが出現している。これらは w* の分布によく対応している（図略）。15 hPa より下では二つの関係が逆転している。

東西風とオゾンのラグ相関を描くと図 3 のようになり、30 hPa 付近では、オゾンの正アノマリーは東西風に約 5 ヶ月先行し、負アノマリーは約 12 ヶ月遅れる。15 hPa より下では逆の関係になり、いずれも観測に基づく解析結果ともよく似ている。

N2O と東西風のラグ相関では N2O がわずか約 1 ヶ月しか先行せず（上空ほど強度が低いので極性はオゾンと反対）、殆ど東西風と同時にあり、また、オゾンの 15 hPa のような位相の変化も存在せず（図略）、パッシブトレーサとして予想される振る舞いである。そうすると、同じ気場に対して、下部成層層でオゾンと N2O が異なる位相を示すことは、オゾンが下部成層層でもパッシブトレーサではなく、ある化学過程が作用していることを示唆している。
新排出シナリオに基づく新しい気候変動シナリオの推計に関する研究
－IPCC AR4 実験－

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1. 背景と目的
気候変動に関する気候推計評価報告書（AR4）作成に向けて、気候モデル（NIES）のアジア太平洋地域統合モデル（AIM）グループの要請にもかなうシナリオ実験を行うことが本研究の目的である。その間にIPCCは全世界での気候変動に対する20世紀と21世紀の気候変動にさらに22世紀まで延長した実験も加えた前例のない種類の一連の実験を行うように要求した。気象研究所では要求のあった全ての実験を行い、これらの実験から得られたデータセットを気候モデル診断相互比較計画（IPCC）においてAR4の第一作業部会の部分を支援しているIPCCモデルアウトプットにデータを提供した。

2. モデル
二酸化硫黄の排出シナリオから硫黄酸化モデルの分布を求めるために、気象研究所で開発された化学輸送モデル（the Model of Aerosol Species IN the Global Atmosphere; MASINGAR; Tanaka et al., 2003）を用いた。MASINGARは大気大循環モデルと四次元データ同化システムに組合されたオンラインの化学輸送モデルである。
地球温暖化予測実験には気象研究所で開発された大気海洋結合モデル（MRI-CGCM2.3: Yukimoto et al., 2003）を用いた。このモデルの大気部分は、水平解像度がT42（格子間隔約270 km）で鉛直30層のハイブリッドσ-p座標系。海洋部分は、水平方向の格子間隔が東西2.5°南北2°で、鉛直23層となっている。赤道付近では南北方向の格子間隔は0.5°となっている。二酸化炭素（CO2）、水蒸気（H2O）、オゾン（O3）、メタン（CH4）、一酸化窒素（NO）を温室効果気体として直接扱っているほか、硫黄酸化モデルによる直接効果を考慮している。

3. 結果
図1に全球年平均の地上気温と降水量の経年変化を示す。予測結果の不確実性を評価するため、結合モデルによる20世紀の気候再現実験（20CM）とIPCCの排出シナリオに関する特別報告書（SRES）のシナリオのうちのA1B、A2、B1については5メンバーのアンサンブル実験を行っており、その個々のメンバーについても線形に示している。温室効果気体の増加に伴い、それぞれのシナリオにおいて地上気温、降水量ともに増加している。年々変動及びアンサンブルメンバー間のばらつきについては、降水量に対してのほうが気温に対してよりも大きくとなっている。1961-1990年の平均に対する2071-2100年における昇温量は、A1B、A1T、A1F1、A2、B1、B2の各シナリオに対してそれぞれ2.4℃、2.1℃、3.2℃、2.7℃、1.7℃、2.0℃となっている。

海面水温の変化量については、ある基準状態からの密度のずれを全球的に積分することに計算することができる。全てのIPCC AR4の実験に対して海面水温の変化を計算した結果を図2に示す。基準状態には太平洋流れの状態におけるカントロール実験の平均的な状態を使用した。2000年から2050年までの間に海面水温は10 cm（SRES B1）から15 cm（SRES A2）の範囲で上昇している。海面水温は二酸化炭素を安定化させた後も上昇を続け、2100年から2300年までの間にさらに10 cm（SRES B1）から15 cm（SRES A1B）上昇している。
図 1 (a) 全球平均地上気温の20世紀の気候再現実験（20C3M）における1961-1990年の平均からの変化（°C）
(b) (a)と同様で全球年平均降水量の変化（%）。アンサンブル実験を行っているものについては個々のメンバーを
綫線で示した。 (a) には比較のため Jones et al. (2001) の観測による結果も黒線で示している

図 2 熱膨張による全球平均海面水位の時間変化（m）。値は産業革命前の状態におけるコントロール実験（Pcontrol）
の最初の50年に対する相対的な値で描かれている。
1. 背景と目的

エルニニョなどに代表される熱帯大気海洋相互作用は、大小さまざまなスケールで生じている。
本研究では、この相互作用の流過程を広い領域かつ分解能が高い数値モデルで直接計算することを目指している。本年度は、モデル中で自発的に生じる地球規模の雲活動の集中化についてしばた。

2. モデル

用いる数値モデルは水平鉛直の2次元であるが、32,768 kmの領域をカバーする積実対流モデルであり、簡単化した雲物理過程を含んでいる。水平解像度は2 km、鉛直解像度は約1 kmであり、一つ一つの雲の内部運動を解像すことができる。海水温度は水平一样である。

3. 成果

図1、図2は時間を追ってモデルの中で雨がどこに降るかを示している。雨の降り方は、モデルの大気に与える放射冷却の構造によって非常に異なる。冷却が下層で強い場合（図1）には雨は全領域で万遍なく降るが、冷却が上層で強い場合（図2）には雨は次第に一ヶ所に集中していく。このような違いは、雲と大規模な大気の流れの相互作用の様相が、放射冷却の仕方で全く異なるために生じる。

現実の熱帯大気での雨の降り方は、普通、海面温度の高低や陸地の分布で決められていると考えられている。しかし、本研究の結果は、条件によっては海面温度などの非一样が始まなくても、雨は自分で集中して降る傾向があることを示している。そのような、強制的な集中化と自発的な集中化の相互作用が、現実大気では重要であるかも知れない。

図1 背の低い放射冷却での降水分布の時間発展。全領域で万遍なく雲活動が生じる。
図2 背の高い放射冷却での降水分布の時間発展。雲活動は一ヵ所に集中していく。
非定常安定成層乱流中のバッシブスカラとアクティブスカラの拡散

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1. はじめに
安定成層乱流中の熱・物質輸送の解明は、その大部分が安定成層流体からなる大気・海洋中の乱流輸送を理解する上で重要である。
しかし、バッシブスカラ（物質）とアクティブスカラ（熱あるいは密度）の拡散の違いに関する基礎研究はきわめて少ない。初期条件と分子拡散係数がバッシブスカラとアクティブスカラでは同じならば、乱流（渦）拡散係数も等しいべきである。そこで、我々のここでの問題は、「いかなる条件下で両者の乱流拡散係数が異なるか？」ということである。
本研究では、非定常安定成層乱流中のバッシブスカラの拡散を、「初期値問題」として考える。
手法としては、線形近似に基づく RDT (Rapid Distortion Theory) 理論を用いる。その結果は、将来の数値計算及び実験における条件設定においても有益であると考えられる。

2. RDT 方程式
鉛直密度勾配が一定で、鉛直物質濃度勾配が一定の流体を考える。支配方程式はプシネスク近似の運動方程式、密度とバッシブスカラ輸送の方程式、それに非圧縮性の条件である。物理量に対するスペクトル分解を支配方程式に代入し、スペクトル成分に対する方程式を得る。さらに非線形項を無視すると、RDT 方程式が得られる。

3. 乱流拡散係数
図 3 は、バッシブスカラの乱流拡散係数 Kc の時間発展を、分子粘性と分子拡散がある場合について示した図である。この図から、Kc は「ゆっくりしたモード」に支配されていることがわかる。

4. 結論
バッシブスカラの乱流拡散係数 Kc においては、初期の密度ゆらぎが小さい場合（PE0/KE0 位置 0.1, ここでは, PE0 は初期の密度ゆらぎによるエネルギー, KE0 は初期の乱流運動エネルギー）でも、「ゆっくりしたモード」が重要になり、この「ゆっくりしたモード」が時間が経過しても減衰しない。本研究の結果と直接比較しうる直接数値シミュレーションはまだ存在しないため、非線形効果を定量的に明らかにするためにも、これらのデータとの比較が今後重要である。また、実際の大気・海洋で存在する風速のシアーの効果の考察が今後必要である。

図 1 バッシブスカラの乱流拡散係数 Kc の時間発展

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赤道域降水活動の構造化に関する水惑星実験とそれに関連する地球型惑星大気大循環モデルへ向けてのソフトウェア開発

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1. 背景と目的

我々の研究の目的は地球の気候状態とその変動を比較惑星科学的見地から考察することである。そのことは同時に地球大気の循環構造を地球流体力学的見地から考察すること、すなわち、現実大気の大循環に内蔵される可能な力学を抽出し、回転成層流体の一般的性のとある実現としてこれを位置づけることでもある。我々の活動は次の二つからなる。第一は、このような目的の追求を可能とするモデル群と関連するソフトウェア群を構築することである。第二は、大気大循環に内蔵された力学の抽出を試みるべく、単純理想的な設定での数値実験を実行することである。境界条件や物理過程などにみられる現実条件での複雑を削除することにより、循環の力学に関する基本的な特異が抽出できることを期待するものである。この意味において我々は一連の実験を基本的実験と称している。これらの基本的実験からは逆に大気モデルを構築する際の指針となる知見の提供が期待される。

2. 赤道域降水パターンに関する水惑星実験

基本的実験の一環として、積雲加熱の鉛直構造が赤道域の降水活動の構造形成に与える影響を調べるため水惑星GCM実験を実行した。その目的は、気候モデルに用いられている程度の空間分解をもつGCMが現実の赤道域の降水活動の可能な形態について理解を深めることにある。実験においては積雲加熱の鉛直分布は赤外放射の吸収係数を変化させることにより変化させた。実験結果はwave-CISK理論の予想とある程度近いものであった。すなわち、積雲加熱のピークが対流層上層に位置すると、モデルの格子点数の降水活動はヒートレンジに組織化され東進する（図1）。降水パターンをキーにしてコンポジット解析を実行した結果も、東進する構造がwave-CISK的であるのに対し、西進する構造は単純な湿潤対流が背景風によって移流されていることを示した（図2）。

3. ソフトウェア開発

上記のようなGCM実験をより簡便に実行すること可能とするソフトウェア環境の構築を目指して、我々はDCPAM、すなわち、階層的な大気モデルと関連するソフトウェア群の開発をすすめている。すでに開発を進めているFortran90をもじいたnetCDFのためのI/Oルーチンを整理統合しここにパッケージ化してgt490ioとしてまとめた。gt490ioを共通のI/O基盤にすえることにより、現在可読性と移植性に配慮した3次元ブリミティブ方程式系モデルへむけてのコード書法の設計を行っている（図3）。

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図 1 赤道域における降水量の経度時間断面図（左）下層冷却実験（右）上層冷却実験

図 2 赤道上における温度場と流れ場の降水に準拠したコンポジット解析図（左）下層冷却実験（右）上層冷却実験

図 3 DMPAM と関連するソフトウェア群
気・液界面直下で発達する対流境界層での乱流混合に関する
3次元直接数値計算

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1. 背景と目的

大気・海洋間で熱・CO₂の交換は地球温暖化・気候システムの変動に対して、メソスケールでは集中豪雨や台風といった異常気象を支配する重要なパラメータである。気・液界面でのこれらの物質交換の駆動力として風によるせん断応力と浮力による対流混合がある。風によるせん断応力による乱流物質交換は理論的手法、室内実験、現地観測、数値計算によって、これまで活発的に進められてきた。しかしながら浮力による乱流混合の影響についてはこれまで議論されず、定量的には未だ不明な点が多い。

Karasudani et al. (1995)は水槽実験により、気・液界面直下での熱対流での熱物質交換係数スセルト数が、固体壁での熱対流よりも2倍近く大きな値を持つことを示した。このことは、東シナ海で観察すると、風速3-4 ms⁻¹程度の風が吹いている時と同等のCO₂の混合効果がある。すなわち海面直下で発達する対流は大気・海洋間での物質交換に大きく寄与しており、この効果を無視することは海面フラックスを大きく過小評価することになる。

本研究の目的は気・液海面直下で発達する乱流熱対流に3次元直接数値計算を適用し、この効果と性質を調べ、Karasudani et al.の結果と比較する。

2. 3次元直接数値計算

冷却された自由表面直下での3次元乱流熱対流に3次元直接数値計算を適用する。計算領域の概略図を図1に示す。せん断力が働いていない自由表面の変動は微少であるため、平坦な滑り壁を自由表面とみなすと仮定する。支配方程式は連続式、ブシネスク近似を用いたNavier-Stokesの式、熱保存式である。

図2に平均温度の初期条件を示す。図2のように初期条件は安定成層状態とする。上部境界での温度(T = -ΔT)を初期条件の上部近傍での温度(T = 0)よりも低い温度とすることにより、計算領域の上部では対流不安定状態となる。さらに初期の温度と流速場にランダムな擾乱を加えることにより乱流状態に発達させる。

境界条件は上部境界条件を滑り条件でT = -ΔT、下部境界条件を滑り無し条件でT = -2ΔTとし、水平方向は周期境界条件を用いた。周期境界条件が大規模流れに影響を与えないように、アスペクト比を6とした。プラントル数は1で固定した。

3. 結果と考察

図1 計算領域の概略図
図2 平均温度の初期分布

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水平方向と時間とで平均化した平均温度の鉛直分布は自由表面直下で線形分布しており、対流不安定の状態であった。レイリー数は 13,771 であった。このレイリー数はレイリー－ベナルデ流のような 2 枚の壁面にはされた流体の対流現象では層流となる。しかし、温度の水平分布等から層流でないことがわかった。

図 3 に鉛直・水平方向の流速の分散、\( \overline{u^2} \)、\( \overline{w^2} \)、の鉛直分布を Karasudani et al. の室内実験の結果と Caughey and Palmer (1979) による固体表面上での気象観測の結果と併せてそれぞれ示す。これらは対流の代表的な速度スケール \( w_* = \sqrt{g \beta F_{wall}} z_i \) で無次元化している。鉛直流速の分散が最大になる高さが室内実験や現地観測に比べて、本計算結果は高いところに位置している。これは DNS でのレイリー数が室内実験や現地観測のものと比べて非常に小さいからである。しかしながら全体的にはい関係を示している。一方、水平方向流速の分散は室内実験結果とは大きさは合っているものの、相関はよくない。室内実験結果では対流混合層の中間 \( z/z_i = 0.5 \) あたりから急激に減少しているのに対し、本計算結果では対流混合層の外縁 \( z/z_i = 1.0 \) あたりで減少し始めている。

ヌッセル数 \( Nu \) とレイリー数 \( Ra \) との関係は一般的に以下の式となる。

\[
Nu = c(Ra)^{1/3}
\]

本計算結果では係数 \( c \) は 0.33 となった。この値は Karasudani et al. の 0.29 と近く、固体表面上での熱対流の 0.15 の 2 倍程度である。固体表面上での自由対流に比べて 2 倍程度の混合効果があることを示している。

図 3 鉛直方向（左）・水平方向流速の分散（右）
準地衡風樞円体渦間の合体法則と樞円体渦モデルの改良

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1. 研究目的
地球流体中では秩序渦構造が安定して長く存在する。McWilliams はスペクトル法による準地衡風乱流の数値計算シミュレーションを実行し、それらの変形・合体が繰り返されることを示した。また、Miyazaki らは樞円体渦モーメントモデルを提案し、2 体渦の Co-Rotating と Counter-Rotating の場合について調べ、渦の合体などの現象を見出した。近年より高いレイノルズ数の計算に対応すべく、CASL (Contour Advective Semi-Lagrangian) 法が Dritschel らにより提案された。本研究では、CASL 法による直接数値計算によってモデルの検証を行いモデルが大いにアスペクト比の正負に拘束される。数値計算においての合体現象の臨界距離を精度よく捉えることを示す。しかし、モデルは異号の「細い渦間の相互作用」において「偽の特性」を予想するため、その改良が必要となる。モデルの改良として離散点近似を導入する。

2. 研究概要と成果
基礎方程式には回転成長流体を有効に近似する f 面上の準地衡風近似の基礎方程式を用いる。初期条件として、2 つの渦を水平方向に a 鉛直方向に h ずらし、アスペクト比を色々変えて CASL 法の計算を実行する。球、扁平樞円体の Co-Rotating の場合 CASL 法による計算とモーメントモデルによる計算はほぼ一致する。しかし、細長い樞円体渦に対してはモーメントモデルの精度が落ちている。

2.1 扁長樞円体渦間の合体現象 (Co-Rotating)
モーメントモデルの計算結果によって引いた合体の確率ラインは、CASL 法の計算結果によって引いた安定とフィラメントの確率ラインに比べると、h>0.5 のときに CASL 法の計算結果より内側にあって、h<0.5 のときは外側にはみ出してずれていることが分かった。

2.2 Counter-Rotating ; ( α/γ =0.3162)
モーメントモデルの計算結果で(図 2)領域(2)では渦は大きく傾き y 軸の負方向に並進する。さらに領域(3)では両方の渦が引き伸ばされ、渦の傾斜角は π/2 に近づき、渦対は y 軸に沿って負方向に移動する。このようなモーメントモデルでは特異な時間発展が起こるのに対し、CASL 法では 2 つの渦が接近するとフィラメントは放出するが、2 体の構造を保ったまま y 軸の正方向に並進運動を続ける。したがって、モーメントモデルの改良が必要となる。

2.3 wire 湧の改良
改良する前の wire 湧モデルで使用されたモーメント近似のわりに、われわれは相互エネルギーやの計算でガウスの線積分法を用いた。その結果、積分の精度が劇的に改善され、2 つの渦が非常に接近しても倒れなくなつった。改良された wire 湧モデルではモーメントモデルで counter-rotating の場合の特異な振る舞いを避ける事ができた。われわれは wire 湧の傾きと太さのパラメータの幾何学関係を正確に提案し、個々の wire 湧が扁長樞円体と仮定する。このようなモデルの改良によって、より太い渦のモデルも精度が上がった。
図 1 では、Co-Rotating の場合、改良モデルの計算によって引いた安定ラインがモーメントモデルの計算結果によって引いた安定ラインよりはるかに CASL 法の計算結果に近寄ったことがわかる。また、Counter-Rotating の場合もその検証を行った。図 2 で示すように、改良する前は 2 つの渦が一番外側のライン内に入ると倒れて計算が止まるようになったが、改良モデルを用いて計算した結果、2 つの渦が重なるほど近づかない限り倒れなくなった。
直線直角座標系における都市スケール大気数値モデルの開発

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1. 背景と目的

数値計算手法の発達と強力なコンピューターの出現により、近い将来、0(〜100m)の水平方向分解能で都市スケールを含む大気地域数値モデルの実行が現実のことになると考えられる。急峻な地形および複雑な物体をより適切に扱うため、適応な数値的手法を提案し、安定、効率的および正確な力学フレームの開発を行う。

2. 概要

我々は、新たな大気局地数値モデルの開発を目的として行った。これは、地球の表面にある急峻な地形および複雑な物体を、より高い分解能で適切に扱うことが期待されるものである。本研究では、非定常三次元可圧縮ナヴィエ・ストークス方程式（Navier-Stokes equations）を計算するためのSIMPLER（Semi-Implicit Method for Pressure-Linked Equation Revised）アルゴリズムとともに、有限体積法（finite volume method）を用いた。我々は、慣例となっている地形追随規格化（terrain-following normalization）は行なわず、高さを垂直座標として用いるデカルト座標を採用した。地球の平均海面高度より上にある急峻な地形および全ての複雑な物体を扱うために、ブロックオフ法（blocking-off method）を導入した。空間および時間に関する離散化については、高次風上対流法（higher-order upwind convection scheme）を採用し、完全時間依存解法（fully time-implicit scheme）を利用してある。ここで乱流計算のテストとして、東京大手町の都市建物群内の流れについて本モデルを実行した。シュミュレーションの結果は、今後の大気局地数値モデル開発において本研究によって提案された数値手法の適用可能性を示すものであった。

3. 成果

図1は、東京大手町の都市建物群の形状及び計算格子を示している。図2に建物群内の地表近くの風速と温度が示されている。図には、物体の角周辺に数値不安定による流れの発生は見られず、シュミュレーションの結果は予想した良いものであったので、この力学フレームに関して有望なものであると言える。今後の課題として都市建物を考慮した放射計算を取り入れる予定である。

図1 東京大手町の都市建物群の形状及び計算格子
図 2 建物群内の地表近くの風と温度分布
ヒートアイランド数値モデルの高分解能化に関する研究

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1. 研究の目的
本研究では、都市キャノピーモデルをメソスケールモデルに組み込むことにより、建物群から地域スケールまでシームレスな数値解析を行う。都市キャノピーモデルを用いることにより建物の高さ方向も含めて気温、風などを予測可能な高分解能モデルの作成を目指す。CFD や風洞実験を活用し都市キャノピーモデルの検証や都市キャノピーモデルを組み込んだメソスケール解析結果をアメダスと比較する予定である。

2. UCSS の開発
都市計画の環境影響評価や政策支援を行うためには都市計画に対応したシミュレーションの入力データの作成やシミュレーション結果を様々な都市情報と一緒に表示・分析を行う必要がある。このような作業を行う上で都市 GIS は極めて有効なツールである。UCSS とは都市気候予測システム(Urban Climate Simulation System)の略称であり、都市気候シミュレーションプログラムを都市 GIS と合わせてシステム化したものである。図 1 に UCSS のソルバー部分の概要を示す。

3. 成果
今年度(利用申請初年度)は、SX-6 上に 3 つの数値モデルのコードを移植し、動作確認及び高速化を実施した。その進行結果を以下に示す。
1) k-ε モデルより单体ブロック周りの流れ場を対象に精度検証を行った。
2) UCSS モデルのベクトル化率の向上及び並列化チューニング・ソース修正を行い、6.9 倍の高速化を達成した。
3) LES モデルのベクトル化率の向上及び並列化チューニング・ソース修正を行い、10.5 倍の高速化を達成した。

図 1 UCSS のソルバー部分の概要

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一様せん断流中の球形液滴に働く抗力と揚力の評価

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1. 背景と目的
流体中での液滴の運動は降雨時や海洋潮流などでの強風時の大気境界層流中でしばしば見られ
る。従って、剪断流中での単一液滴の飛散機構を解明することは、降雨や砂波時に海面から飛散
する液滴が、大気・海洋間での物質輸送に及ぼす効果を論じるうえで極めて重要になる。

そこで本研究では、一様せん断流中の単一球形液滴周りおよび液滴内の流れを三次元直接数値計
算法(Direct Numerical Simulation;DNS)を初めて適用することにより、球形液滴に働く抗力と揚力
に及ぼす流体中の速度こう配の効果を明らかにすることを目的とした。

2. 概要
単一球形液滴周りおよび液滴内に対して直接数値計算法を適用することで、球形液滴周りおよ
び液滴内の流動場を明らかにし、その結果を用いて液滴に作用する抗力および揚力を評価した。
計算の概略を図1に示す。周囲流を一様せん断流とし、液滴表面で与えられた境界条件は、1) 流
体は界面を横切らず、液滴は変形しないこと、2) 液滴表面に対する接線方向のせん断応力が連続
性を満たすこと、3) 液滴表面での接線方向の速度を連続性を満たすこと、であった。DNSを実行
する際のパラメータは球周りでのレイノルズ数$Re_x$および無次元速度こう配$\alpha$であり、$\alpha=0.0$, 0.1,
0.2, 0.3, 0.4において$Re_x=1\sim300 (1, 5, 10, 50, 100, 300)$の範囲で計算を実行した。

3. 成果
図2に一様流中($\alpha=0.0$)における液滴の抗力係数$C_D$のレイノルズ数$Re_x$に対する変化を示す。図2
中の実線はBeard & Pruppacher (1969)の実験関連式、破線はGunn & Kinzer (1949)の実験関連
式である。図から本数値計算の結果は実験値と良好に一致していることがわかる。図3に周囲流の無次
元速度こう配$\alpha=0.1, 0.2, 0.3, 0.4$での抗力係数$C_D$と一様流中($\alpha=0.0$)における液滴の抗力係
数$C_D$との比のレイノルズ数$Re_x$に対する変化を示す。図から抗力係数は$\alpha$の増加に伴い、増加す
ることがわかる。また、その影響は$Re_x$の増加に伴って大きくなることがわかる。図4に液滴に
作用する揚力係数$C_L$のレイノルズ数$Re_x$に対する変化を示す。図から、一様流中における液滴に
働く揚力は常に零であることがわかる。また、液滴に働く揚力は速度こう配の存在する場合には
$Re_x$の増加に伴って減少し、零となり、やがて負の値となることがある。さらに、$\alpha$の増加に伴
い、周囲流の速度こう配による揚力への影響が大きくなることがわかる。この揚力のレイノルズ数
に対する変化はKurose & Komori(1999)により得られた剛体球に働く揚力と同様の傾向を示す。
図 2 一様流中の液滴に作用する抗力係数 $C_D$

図 3 一様せん断流中の液滴に作用する抗力係数 $C_D$

図 4 一様せん断流中の液滴に作用する揚力係数 $C_L$
表面侵食モデルによる嘉陵江流域の土砂生産量推定
－長江（揚子江）上流域を対象として－

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1. 背景と目的
三峡ダム(Three Gorges Dam)の洪水が2003年6月より開始され、ダム湖では、既に流入する微細土砂が沈降し、流入水の増加に比べて湖内水が濃んでいることが報告されている。長江上流全域での土砂生産域のうち、図1に示す嘉陵江流域の特定地域7万km²で三峡ダム地点の通過量の43%に及ぶ土砂が生産されていると推測されている。従って、こうした集中生産地区での土砂動態を検討することが、三峡ダム湖への流入する土砂の管理を進める上で、必要となる。本報では、三峡ダム湖への土砂連続量に占める割合が高い嘉陵江流域での土砂動態を明らかにする第一歩として、表面侵食を土砂生産の主要機構とするモデルの適用を試みた。適用したモデルは、(1)侵食の主要外力を降雨衝撃とするUUSLE(Universal Soil Loss Equation)を表面流侵食型に修正した日下・藤田、木本らのモデル、(2)著者らが提案するunit stream power形式の表面流侵食モデル、(3)土砂供給源としての河岸を重要視した氾濫堆積土砂の侵食モデルの3つのモデルである。対象年度としては1987、1988年とした。

2. 表面土壌侵食モデルの嘉陵江流域への適用
(1) 降雨流出モデルの適用
適用するモデルの外力強度(山腹斜面上の表面流、河川流量)の推定のため、流出水文モデルとしてStanford Watershed Modelを採用した。計算結果は、土壌地表の浸透能が高く、短期間の集中的な降雨時以外、表面流は発生しにくいことを示し、長江上流の流域斜面での土砂生産は降雨期間を集中するという特徴に対応した結果が得られている。
(2) 表面土壌侵食モデルの適用結果
1987年の嘉陵江流域での土砂生産量推定に用いられた3種のモデル中のパラメータを適宜決定した適用結果が、図2、3、4である。
図2は年継続流出土砂量の観測値と、各モデルによる計算値とを比較したものので、侵食外力を降雨流と河川流、雨滴侵食を主要外力とする流れのモデルでも、年単位の生産量推定は可能ですことを示している。
図3は月別流出土砂の観測値と計算値とを比較したものである。斜面由来の土砂生産量の推定においては、表面流水外力を有するモデルでは表面流を発生させる強い降雨が集中する降雨期にのみ土砂生産が行われるのに対し、雨滴侵食モデルでは年間を通じて生産が行われることになる。しかしながら、両モデルとも観測値の月別傾向を十分には明確しているとは言い難く、年間を通じての変異を再現するためには、生産源としての河岸の取り扱いが重要であることを示している。河岸堆積土砂侵食モデルは、年間の月別変動を概ね再現している。
図4は日単位の流出土砂の観測値と計算値とを比較したものので、変動パターンは再現できるものの、量的な推定には至っていないことがわかります。
以上の計算結果から、嘉陵江流域ではmass movementの発生領域区分に関わらず、表面侵食モデルで年単位および月単位では、流出土砂量の推定はある程度可能と考えられる。ただし、月単位の推定精度の向上と日単位の変動までの再現のためには、1) 斜面生産土砂の河岸に至るまでの連続性と時間的な遅れを考慮した斜面からの供給タイミングについてのモデル化、2) 非定常性が強い土砂生産を如何に取り込むか、の検討が必要と考えられる。

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図 1 嘉陵江の位置

図 2 年生産量の計算値と観測値の比較

図 3 月別生産量の計算値と観測値の比較

図 4 時間変動の計算値と観測値の比較
森林の光環境シミュレーション

武田知己・小熊宏之
国立環境研究所地球環境研究センター

1. はじめに
リモートセンシングは、森林に入射した日射が森林内で吸収・散乱され結果、森林の生理状態に応じて量的・質的な変化を受けることを利用し、森林の波長別反射特性を測定して森林の生理状態を推定している。しかし、森林群落による日射の波長別反射率は、森林の生理状態だけでなく、森林の構造にも大きく影響を受けることから、リモートセンシングによって得られるデータを正しく理解するためには、森林の構造を明らかにする必要である。そこで本報告書では、レーザスキャナを用いた森林の三次元構造の測定方法とその結果、および森林の三次元構造を用いた光環境のシミュレーション方法とその結果について述べる。

2. 方法
2003年4月から12月に、北海道苫小牧市内の国有林内にある苫小牧Flux Research Site（42°44′N、141°31′E）（以下、苫小牧FRS）において、ニホンカラマツ（Larix kaempferi Sarg.）を対象に実験を行った。苫小牧FRSのニホンカラマツは、樹齢が約47年、平均樹高が18m、植栽密度が681.4本/haだった。

森林の構造を表すパラメータとして、単位地面積あたりの葉面積であるLAI（Leaf Area Index）と平均葉傾斜角であるALIA（Average Leaf Inclination Angle）を用いた。森林の構造を三次元で表すため、40m×40mの測定区画を8m×8mの25区画に分割し、各区画を垂直方向に1mの水平な層で分割して得られるセルに対してLAIとALIAを測定した。LAIとALIAの測定にはレーザスキャナを使用し、セルに入射するレーザ光線の数とセルを通じるレーザ光線の数の比からGap Fractionを測定した。Gap FractionからLAIとALIAを計算する方法には、Norman and Campbell（1989）の方法を用いた。

森林に入射した日射の吸収・散乱過程をシミュレーションするために、コンピュータグラフィックの技術であるレイティングを用いた。レイティング法で必要となる構造パラメータは、レーザスキャナによる測定から求めたLAIとALIAを、光学パラメータは、2003年8月に測定した日射の散乱係数を用いた。

3. 結果
図1に、同じ高さのLAIを25区画で平均したカラマツ林の垂直分布と標準偏差を示す。LAIの垂直分布は、2.5mと10.5mにピークを2つ持っている。10.5mのピークはカラマツ、2.5mのピークは下層木の影響と考えられる。

図2に、測定で求めた個葉の散乱係数とシミュレーションで求めた群落の反射特性を波長別に示す。

群落の分光反射特性は個葉とほとんど同じ変化のパターンを示したが、絶対値をみるとかなり小さくなる事が分かった。

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図 1 2003年8月に測定した葉の垂直分布
図 2 個葉の分光散乱係数（点線）と群落の分光反射特性
Program of
the 12th Supercomputer Workshop

October 4, 2004
National Institute for Environmental Studies
Tsukuba, Ibaraki, Japan
Program of The 12th Supercomputer Workshop

11:00～17:25, October 4, 2004
National Institute for Environmental Studies, Tsukuba, Japan

11:00～11:05 Opening address
Shuzo Nishioka (NIES)

11:05～11:15 Introduction of research programs
Yasumi Fujinuma (CGER/NIES)

11:15～11:30 Analysis of year-to-year ozone variation over the subtropical Western Pacific region using EP-Toms data and CCSR/NIES nudging CTM
Hideharu Akiyoshi (NIES), et al.

11:30～11:45 Ozone QBO simulated with MJ98-CTM
Kiyotaka Shibata (Meteorological Research Institute), et al.

11:45～12:00 Simulation of spectral reflectance of forest
Tomomi Takeda (CGER/NIES), et al.

12:00～13:00 Lunch

13:00～13:15 Role of groundwater on water cycle and nutrients inventory (Supply) in shallow eutrophic lake, Kasumigaura, Japan
Tadanobu Nakayama (NIES), et al.

13:15～13:30 Impact of environmental change in Changjiang River catchment on marine ecosystem in the East China Sea
Masataka Watanabe (NIES)

13:30～13:45 The effects of dispersing droplets on the mass transfer across the air-sea interface
Satoru Komori (Kyoto University)

13:45～14:00 Numerical simulation of the airflow within and above urban canopies using the spatially averaged k-ε Model
Yasunobu Ashie (Building Research Institute), et al.

14:00～14:15 Merger of Quasi-Geostrophic ellipsoidal vortices and refinements on the ellipsoidal moment
Eita Ri (University of Electro-Communication), et al.

14:15～14:30 Heat and mass transfer in stratified turbulence
Hideshi Hanazaki (Kyoto University)

14:30～14:45 Representation of complex objects in a new Cartesian Coordinate urban street atmospheric numerical model
Weiming Sha (Tohoku University)

14:45～15:00 Formation of eddies behind a large mountain in stratified rotating flow
Yu Hozumi (Kyoto University), et al.

15:00～15:10 Coffee Break
15:10～15:25 Numerical experiment on the large-scale organization of cumulus convection
Kensuke Nakajima (Kyushu University)

15:25～15:40 Dependence of tropical precipitation activities on the vertical of radiative cooling observed in an aqua planet simple structure GCM
Yoshi-Yuki Hayashi (Hokkaido University), et al.

15:40～15:55 Study of the estimate of New Climate Change Scenarios based on New Emission Scenarios
Takao Uchiyama (Meteorological Research Institute), et al.

15:55～16:10 Synoptic scale variability of atmospheric CO₂ in continental boundary layer: model and observations
Shamil Maksyutov (Frontier Research Center for Global Change), et al.

16:10～16:25 Development of dynamic-vegetation coupled AGCM
Ryuta Oishi (Center for Climate System Research, University of Tokyo), et al.

16:25～16:40 Contribution of land surface processes to precipitation variability in an atmospheric General Circulation Model
Tomohito Yamada (Institute of Industrial Science, University of Tokyo), et al.

16:40～16:55 Reproducibility of climate change in the 20th century with a coupled ocean-atmosphere GCM
Toru Nozawa (NIES), et al.

16:55～17:05 The flow of program tuning of a supercomputer systems
Kumiyasu Hamada (NEC)

17:05～17:20 Discussion

17:20～17:25 Closing address
Gen Inoue (CGER/NIES)
スーパークンピュータによる地球環境研究発表会（第12回）プログラム

日時：平成16年10月4日（月）11:00～17:25
場所：独立行政法人国立環境研究所地球温暖化研究棟 交流会議室

11:00～11:05 開会挨拶
西岡 秀三（国立環境研究所地球環境研究センター長）

11:05～11:15 スーパークンピュータ利用研究概要紹介
藤沼 康実（国立環境研究所地球環境研究センター研究管理官）

11:15～11:30 EP-TOMSデータとCCSR/NIESナッシャング化学輸送モデルを用いた亜熱帯西太平洋海域オゾンの年々変動の解析
秋吉 莉治1・L.B. Zhou2・高橋 正明3（1国立環境研究所成層層オゾン層変動研究プロジェクト、2中国科学院大気物理研究所、3東京大学気候システム研究センター）

11:30～11:45 気象研究所成層層化学輸送モデルによるオゾンQBOについて
柴田 清孝・出牛 真（気象研究所 環境・応用気象研究部）

11:45～12:00 森林の分光反射特性のシミュレーション
武田 知己・小熊 宏之（国立環境研究所地球環境研究センター）

12:00～13:00 昼休み

13:00～13:15 NICE（NIES Integrated Catchment-based Eco-hydrology）モデルの使用による霞ヶ浦流域での地下水流量が水循環に及ぼす影響の再現シミュレーション
中山 忠晴1・渡辺 正孝2（1国立環境研究所流域環境管理研究プロジェクト、2国立環境研究所水士壌環境研究領域）

13:15～13:30 長江流域の環境変動が東シナ海の海洋生態系に与える影響
渡辺 正孝（国立環境研究所水士壌環境研究領域）

13:30～13:45 大気海洋間の物質輸送に及ぼす飛散液滴の効果
小森 悟（京都大学大学院工学研究科）

13:45～14:00 h-h空間平均モデルを用いた都市キャノピー内及び上空の気流解析
足永 靖信1・一ノ瀬 俊明2・河野 孝昭1（1建築研究所環境研究グループ、2国立環境研究所地球環境研究センター）

14:00～14:15 準地衡風積層体地形の合体法則と積層体地形モデルの改良
李 英太・平 寛史・宮嶋 武（電気通信大学大学院）

14:15～14:30 成層乱流中の熱・物質輸送
花崎 秀史（京都大学大学院工学研究科）

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14:30〜14:45  代表作:Complex Objects in a New Cartesian Coordinate Urban Street Atmospheric Numerical Model (直角直線座標系における大気数値モデル中の地形・物体の表現について)
○余 偉明 (東北大学大学院理学研究科)

14:45〜15:00  大規模山岳風下側における渦の鉛直構造
○細谷 祐1・種田 洋匡3・余 偉明2 (1京都大学防災研究所、2東北大学大学院理学研究科)

15:00〜15:10  休憩

15:10〜15:25  積雲対流の大規模組織化についての数値実験
○中島 健介 (九州大学大学院理学研究院)

15:25〜15:40  簡略な水蒸気 GCM に見られる熱帯域降水活動の放射冷却率鉛直分布依存性
○林 祥介1・石渡 正樹2・小高 正嗣1・山田 由貴子1・中島 健介3 (1北海道大学大学院理学研究科、2北海道大学大学院地球環境科学研究院、3九州大学大学院理学研究院)

15:40〜15:55  新排出シナリオに基づく新しい気候変動シナリオの推計に関する研究
○内山 貴雄・楠 昌司・行本 誠史 (気象研究所 気候研究部)

15:55〜16:10  Synoptic Scale Variability of Atmospheric CO2 in Continental Boundary Layer: Model and Observations (大陸大気境界内の CO2 濃度のシノプティックスケール変動: 手法と観測)
○Shamil Maksyutov1・Misa Ishizawa1・Gen Inoue2 (1Frontier Research Center for Global Change, 2Center for Global Environmental Research/National Institute for Environmental Studies)

16:10〜16:25  動態植生結合大気大循環モデルの開発
○大石 龍太・阿部 彩子 (東京大学気候システム研究センター)

16:25〜16:40  大気大循環モデルを用いた降水変動に対する陸面影響評価
○山田 朋人1・森 信次郎2・沖 大幹1 (1東京大学生産技術研究所、2総合地球環境学研究所)

16:40〜16:55  大気海洋結合モデルによる 20 世紀の気候再現性
○野沢 徹1・水島 達也1・小倉 知夫1・横倉 徳夫1・岡田 直資1・塩竈 秀夫1・江守 正多12・阿部 彩子23・矢尾 博康3・木本 昌秀3 (1国立環境研究所大気圈環境研究領域、2地球環境フロンティア研究センター、3東京大学気候システム研究センター)

16:55〜17:05  スーパーコンピュータのプログラムチューニングの流れ
○矢田 邦靖 (NEC/HPC グループ/HPC センター)

17:05〜17:20  総合討論

17:20〜17:25  閉会挨拶 井上 元 (国立環境研究所地球環境研究センター総括研究管理官)

（○印は発表者です）
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