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Foreword

The Center for Global Environmental Research (CGER) at the National Institute for Environmental Studies (NIES) was established in October 1990. CGER’s main objectives are to contribute to the scientific understanding of global change and to identify solutions for pressing environmental problems. CGER conducts environmental research from interdisciplinary, multi-agency, and international perspectives, provides an intellectual infrastructure for research activities in the form of databases and a supercomputer system, and makes the data from its long-term monitoring of the global environment available to the public.

CGER installed its first supercomputer system (NEC SX-3, Model 14) in March 1992. That system was subsequently upgraded to an NEC Model SX-4/32 in 1997, and since 2002 we have been using an NEC Model SX-6. The system is currently working at its full capacity due to the increased use of the CPU and we are looking into ways to improve its effectiveness.

The Supercomputer Steering Committee evaluates proposals of research requiring the use of the supercomputer system. Proposals can be submitted by researchers from NIES and other research organizations including universities in Japan. The committee consists of leading Japanese scientists in climate modeling, atmospheric chemistry, ocean environment, computer science, and other areas of concern in global environmental research. In the 2005 fiscal year (April 2005 to March 2006), sixteen proposals were approved.

Supercomputer Activity Report Vol.14 is a record of research results obtained in the 2005 fiscal year. Research papers are classified into four categories—Climate Modeling, Atmospheric and Oceanic Environment Modeling, Geophysical Fluid Dynamics, and Other Research—together with an overview of the supercomputer system. It is important to note that the papers in this report do not necessarily reflect the final results of the research; the final results are reported in full papers upon completion of the individual research programs.

We hope this report provides useful information on global environmental research, and we look forward to your suggestions and comments on the use of supercomputers in our research.

January 2007

Yasuhiro Sasano
Director
Center for Global Environmental Research
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1. Climate Modeling
A Future Ozone Layer Prediction Using CCSR/NIES Chemical Climate Model with T42 Horizontal Resolution

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Abstract

A preliminary result of a future ozone prediction using Center for Climate System Research/National Institute for Environmental Studies (CCSR/NIES) Chemical Climate Model is presented. The calculation was performed under the future projection of greenhouse gases and ozone depleting substances. The IPCC A1B scenario and the World Meteorological Organization (WMO) A2 scenario are used for the future greenhouse gases and halogens, respectively. The result shows that Antarctic ozone hole will disappear by the middle of this century according to a decrease in the chlorine loading in the Antarctic lower stratosphere, although there are year-to-year variations in ozone hole reflecting the dynamics of the Antarctic polar vortex. The result indicates the importance of the chlorine decrease in the atmosphere to the ozone hole recovery.

Keywords: Ozone hole, Chemical Climate Model, Chlorine loading, Polar vortex, Stratosphere

1. Introduction

Chemical Climate Model (CCM) is a useful tool to predict future ozone layer and examine the mechanism for ozone variations in the future. CCM includes dynamical, radiative, and chemical processes in the atmosphere and also includes interaction processes between them. Since Antarctic ozone hole is a phenomenon in the polar region where highly non-linear processes of the polar vortex is dominant, a three-dimensional numerical model coupled with chemistry is needed to simulate it.

Nagashima et al. (2002) developed a CCM for future ozone layer prediction. The CCM include interaction processes between dynamics, radiation, and chemistry. The horizontal resolution is T21 (5.6° by 5.6°). They demonstrated that the Antarctic ozone hole will disappear around 2040. One of defects of this CCM is that Antarctic ozone hole occurs in November, about one month later than observations, with insufficient ozone depletion in October. The late maturity of ozone depletion results from unusually late Antarctic polar vortex breakup in the model and an insufficient chemical ozone destruction. The late vortex breakup is due to a too stable Antarctic polar vortex in the winter and spring of the model. In order to improve these defects for ozone hole predictions, we made several improvements in the model. We improved horizontal resolution to T42 (2.8° by 2.8°) and introduced a new chemical scheme including bromine chemistry and heterogeneous reactions on Polar
Stratospheric Clouds (PSCs) of Super-cooled Ternary Solution (STS) as well as Nitric Acid Trihydrate (NAT) and ice particles. We also included radiation process of Schumann-Runge bands, effects of atmospheric sphericity in solar radiation, non-orographic gravity wave effects, and sedimentation velocity corresponding to PSC radius. The model was run imposed by a future projection of greenhouse gases and halogens from the IPCC A1B scenario and Ab scenario of the Beijing Amendments. The results were analyzed and investigated in terms of the recovery of ozone hole in the future atmosphere.

2. Model Descriptions and Numerical Experiments

CCSR/NIES CCM was developed based on the version 5.4g of CCSR/NIES AGCM (Nagashima et al., 2002; Takigawa et al., 1999). The model has been improved and updated since the last future ozone prediction calculation (Nagashima et al., 2002) in the following respects.

Chemistry:

Heterogeneous reactions on PSCs of Supercooled Ternary Solution are incorporated. The scheme was originally developed by Sessler et al. (1996) for chemical box model and is modified for the CCSR/NIES CCM.

Gas phase and heterogeneous reactions related to bromine species are incorporated. The scheme was developed by Akiyoshi (2000) in a frame of the family method.

In Nagashima et al. (2002), constant sedimentation velocities of PSCs were assumed; 1 ms\(^{-1}\) for NAT particles and 10 ms\(^{-1}\) for ice particles. In the new CCM, the number density of PSCs for a log-normal distribution with the dispersion of 1.8 is assumed. The specified number densities of STS, NAT, and ice particles are 10 particles cm\(^{-3}\), 0.005 particles cm\(^{-3}\), and 0.005 particles cm\(^{-3}\), respectively. The number densities of NAT and ice are based on the observation by Waibel et al. (1999). The mode radius is calculated by the condensed mass assuming sphericity of the particles. Then sedimentation velocity is calculated using the mode radius.

Radiation:

Since CCSR/NIES AGCM does not include radiation process of wavelength less than 200 nm, a scheme for photolysis rate calculation in the Schuman-Runge bands (177.5 nm-202.5 nm) is developed using the parameterization of Minschwaner et al. (1993), and incorporated. The Schumann-Runge bands photolysis rates of 27 species are calculated (O\(_2\), O, N\(_2\)O, H\(_2\)O, NO\(_2\), HNO\(_3\), HNO\(_4\), N\(_2\)O\(_5\), ClONO\(_2\), CCl\(_4\), CFC\(_3\)(CFC-11), C\(_2\)Cl\(_2\)(CFC-12), CH\(_3\)Cl, HCl, HOCl, CH\(_3\)CCl\(_3\), Cl\(_2\)O, CF\(_2\)ClCFC\(_2\)(CFC-113), HBr, BrONO\(_2\), BrCl, CHBr\(_3\), CH\(_2\)Br, CF\(_2\)ClBr(Halon-1211), CF\(_3\)Br(Halon-1301), CHF\(_2\)Cl(HCFC-22), and OCS). This improved the vertical profile of N\(_2\)O (Akiyoshi et al. 2002a, b), CFCs, and halons.

A radiation calculation scheme that includes effects of atmospheric sphericity was developed by a CCSR/NIES group (Kurokawa et al., 2005). The plane-parallel radiation transfer scheme of CCSR/NIES AGCM was modified by a pseudo-spherical approximation. The new scheme shifts the onset time of ozone hole 10 days earlier than that from the plane parallel scheme and the result corresponds more with observations.
Dynamics:
The horizontal resolution is changed from T21 to T42.

In order to include effects of sub-grid scale, McFarlane (1987)'s parameterization is used for orographic gravity wave drag calculations. For nonorographic gravity wave effect, a parameterization by Hines (1997) is incorporated into the CCM. These schemes calculate gravity wave drag to the zonal wind at the CCM grids. Values of parameters for Hine's parameterization are as follows: A Gaussian zonal wind amplitude of 1.5 ms$^{-1}$ around the equator with the half width of 15 degrees is assumed and added to a constant value of 1.5 ms$^{-1}$.

3. Numerical Experiments

Numerical experiment for future ozone layer prediction is performed following the REF2 scenario of Chemical Climate Model Validation (CCMVal). CCMVal is a project under Stratospheric Processes and their Role in Climate (SPARC) of the World Climate Research Programme (WCRP). The goal of CCMVal is to improve understanding of CCMs and their underlying General Circulation Models (GCMs) through process-oriented validation. REF2 scenario is one of the scenarios for future ozone prediction. In this scenario, the surface time evolution of CO$_2$, CH$_4$, and N$_2$O is taken from the IPCC (2000) A1B scenario. That for halogen gases is from the Ab scenario in WMO (2003). Fig. 1 shows the time evolutions of CO$_2$, CH$_4$, N$_2$O, and CCl$_4$ at the surface of the model. CCl$_4$ is a total chlorine that is included in source gases (CCl$_4$, CFC$_3$(CFC-11), CF$_3$Cl(CFC-12), CH$_3$Cl, CH$_3$CCl$_3$, CF$_2$ClCFCl$_2$(CFC-113), CHF$_2$Cl(HCFC-22), and CF$_2$ClBr(Halon-1211) for this model) The CCl$_4$ at the surface increased rapidly in the 1980s and reached a maximum around 1995 and then decrease to the future slowly. The output of sea surface temperature (SST) from a numerical experiment on an atmosphere-ocean coupled general circulation model (CGCM) called The Model for Interdisciplinary Research on Climate (MIROC, Hasumi and Emori, 2004, Shiogama et al., 2005) are used as the SST data in the CCM. MIROC was developed by CCSR, NIES, and the Frontier Research Center for Global Change (FRCGC).

![Fig. 1. Time evolution of CO$_2$, CH$_4$, N$_2$O, and CCl$_4$ at the surface of Chemical Climate Model. Concentrations of CO$_2$ and N$_2$O presented are multiplied by a factor of 10.](image-url)
The effect of QBO, the solar 11 year cycle, and enhancement of stratospheric aerosols due to volcanic eruptions are excluded.

The initial profiles of temperature, winds, and chemical species are taken from an output of a spin-up run with 1975 conditions. The calculation starts on 1 January 1975 and ends on 1 January 2051. The data from 1 January 1980 are used for analysis.

4. Results

The T42 CCM developed in this study simulates October minimum of total ozone in a seasonal cycle, while the previous CCM used in Nagashima et al. (2002) had the minimum in November. The improvements in chemical and dynamical processes lead to this better simulation of seasonal cycle of Antarctic ozone.

Fig. 2 shows time series of ozone hole area defined by the area where total ozone amount is less than 220 DU in the south of 40°S. TOMS observation shows appearance of ozone hole at the beginning of 1980s followed by the rapid growth in 1980s, the slower growth in 1990s, and a maximum of ozone hole area around 2000. The CCM result simulates the time evolution very well. In particular the appearance of ozone hole at the beginning of 1980s and a maximum around 2000 are reproduced well. CCM shows Antarctic ozone in the future atmosphere: Ozone hole will not be reduced until 2020. A clear reduction will be seen after 2020 and ozone hole will be disappeared in the middle of this century. The data around 2050 indicates that ozone hole does not totally disappear, because there is an year-to-year variation in the dynamics of the atmosphere and ozone amount over Antarctica is sensitive to the Antarctic polar vortex condition.

Fig. 3 shows time evolution of minimum total ozone in the south of 40°S. A time evolution reversed to that of ozone hole area in Fig. 2 is seen. The minimum total ozone decrease rapidly in the 1980s, more slowly in 1990s, reached a minimum around 2000, then increased and recovered in the middle of this century.

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Fig. 2. Time evolution of ozone hole area. Black squares represent observation by TOMS instrument and red circles represent the model results of CCMVal REF2 simulation.

Fig. 3. Time evolution of total ozone minimum in the south of 40oS. Black squares and red circles are same as Fig. 2.
Fig. 4 shows time evolution of the amount of total reactive chlorine (Cl\textsubscript{y}) at 50 hPa averaged at 80\textdegree S. The Cl\textsubscript{y} amount reached a maximum around 2000, about five year later than the maximum of CCl\textsubscript{y} at the surface. The time evolution of Cl\textsubscript{y} corresponds well to those of the ozone hole area and the minimum total ozone. This indicates that the time evolution of Cl\textsubscript{y} in the Antarctic lower stratosphere is a key factor to that of ozone hole. Satellite observation of HCl indicates that Cl\textsubscript{y} was about 3.3 ppb in 2005. The peak value of Cl\textsubscript{y} in the model is a little smaller than the observation.

**Fig. 4.** Time evolution of zonal-mean total reactive chlorine at 80\textdegree S and 50 hPa in Chemical Climate Model.

Fig. 5. Polar map of total ozone amount in October. (a) 1979-1982 average of TOMS, (b) 1989-1992 average of TOMS, (c) 1998-2001 average of TOMS, (d) 1978-1982 average of CCM, (e) 1988-1992 average of CCM, and (f) 1998-2002 average of CCM. Projection is stereographic, 180\textdegree E at top and 90\textdegree W at left. Color scale in Dobson unit is presented at the right bottom.
Polar map of October mean TOMS total ozone averaged 4 years are presented in the upper of Fig. 5. These maps show that ozone hole developed over Antarctica from 1980 to 2000. The lower column presents CCM results. It is obvious that the model simulates the evolution of ozone hole during this period very well.

Fig. 6 shows October mean total ozone simulated by CCM after 2010. Five year averages centered at 2010, 2020, 2030, and 2040 are shown. Ozone hole is reduced apparently after 2020.

Fig. 6. Same as Fig. 5, but all from CCM. (g) 2008-2012 average, (h) 2018-2022 average, (i) 2028-2032 average, and (j) 2038-2042 average. Color scale is also same as Fig. 4.

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References


**Publications and Presentations**

**Original Papers and Reviews:**


**Conference Reports:**


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Influence of Dynamic Vegetation Change on Climate Change 
Due to Increase of CO₂

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Abstract
A dynamic global vegetation model (DGVM) is coupled to an atmospheric general 
circulation model (AGCM) in order to investigate the role of vegetation dynamics in 
the global warming. In order to investigate the role of the dynamic vegetation change 
due to high CO₂ and the consequent change of temperature, precipitation and carbon 
budget, the equilibrium state of climate and biospheric properties which are induced 
by DGVM-AGCM experiments and fixed vegetation experiment which does not adopt 
the DGVM are compared between preindustrial, doubled and quadrupled atmospheric CO₂ values (285 ppm, 570 ppm and 1141 ppm). Results show that global 
warming is amplified due to the inclusion of dynamic vegetation. This is because the 
high CO₂ and consequent climate change increase net primary productivity (NPP) 
and vegetation coverage. Result also shows that terrestrial carbon storage decreases 
due to the inclusion of dynamic vegetation. The major cause is acceleration of soil 
carbon decomposition due to amplification of warming. Overall, the inclusion of 
dynamic vegetation process enhances the warming compared to the fixed vegetation 
experiments under the high CO₂.

Keywords: AGCM, Dynamic vegetation, Global warming

1. Introduction

Anthropogenic increase in atmospheric concentration of CO₂ and other greenhouse gases 
is predicted to cause significant warming of climate. This change of CO₂ and warming 
considerably increases net primary productivity (Mellilo et al., 1993), change vegetation 
distribution (Cramer et al., 2001) and terrestrial carbon storage (Gerber et al., 2004). These 
processes are incorporated into general circulation model, and a number of global warming 
experiments are performed. In one of these global carbon cycle experiments under future 
global warming scenario, strong positive feedback from terrestrial biosphere to atmosphere 
was noted (Cox et al., 2000). On the other hand, another study did not show this positive 
feedback (Dufresne et al., 2002). Recent studies still show this discrepancy in future 
terrestrial carbon projection in C4MIP (Friedlingstein et al., 2006). These models used in 
C4MIP are combination of different atmosphere models, ocean models, land surface models, 
ocean biosphere models and terrestrial biosphere models. The resultant change of terrestrial 
carbon storage reflects all these model uncertainties. Furthermore, some of terrestrial 
biosphere models are able to predict dynamical change of vegetation distribution but others 
assume no vegetation change under global warming. However, there are no former researches 
which investigate the influence of dynamic vegetation change to atmosphere compared to 
fixed vegetation.

In the present study, we coupled an AGCM and a DGVM in order to investigate the 
influence of dynamic vegetation under high CO₂. We performed control experiment and
warming experiment with quadrupled atmospheric CO₂ using this coupled model and usual AGCM with fixed vegetation. Result of the present study helps to understand the discrepancy among former model experiments.

2. Model and Experimental Design

2.1 Model

In the present study, Lund-Potsdam-Jena DGVM (LPJ-DGVM; Sitch et al., 2003) is coupled to CCSR/NIES/FRCGC Model for Interdisciplinary Research on Climate (MIROC; Hasumi and Emori, 2004). LPJ-DGVM includes processes based on a large-scale terrestrial carbon exchange module and a vegetation dynamics module. In this model, vegetation types are represented by 10 plant functional types (PFTs) which can coexist with other PFTs if conditions and competition allow. The LPG-DGVM obtains monthly mean temperature, precipitation and cloud cover ratio which corresponds to the solar irradiance from MIROC. These atmospheric variables are interpolated linearly to daily value and photosynthesis, carbon allocation, vegetation growth, competition, death and establishment are calculated for each PFTs every year. In each land grid cells, the combination of PFTs is translated into vegetation type adopted in land surface model Minimal Advanced Treatments of Surface Interaction and Runoff (MATSIRO; Takata et al., 2003) which is part of MIROC. LPJ-DGVM is also able to predict land carbon storage and carbon exchange, but carbon feedback to atmosphere is not adopted in the present study. Fig. 1 shows the summarized flowchart of this LPJ-MIROC developed in the present study. This model has shown a general agreement with the present day climate and vegetation distribution.

Fig. 1. Flowchart of LPJ-MIROC.
2.2 Experimental design

In this experiment, horizontal and vertical resolutions of atmospheric grid points are adopted to T42 vertically and 20 layers horizontally, respectively. Experimental setup is the same as MIROC3.2-midresolution (Hasumi and Emori, 2004) submitted to the Intergovernmental Panel on Climate Change assessment report 4 (IPCC AR4) and coupled carbon cycle climate model intercomparison project (C4MIP, Friedlingstein et al., 2006) except for the inclusion of DGVM. The atmospheric CO₂ is set to 285 ppm in experiment CTRL, 570 ppm in experiment 2xCO₂ and 1141 ppm in experiment 4xCO₂, respectively. Sea surface temperature and sea ice are prognosed by slab ocean model. Other parameters (e.g. orbital element) are set to those of the present climate. 150 years integrations are performed for these experiments to reach equilibria, using no-vegetation initial condition. Additionally, experiments named CTRLfix, 2xCO₂fix and 4xCO₂fix are performed to identify the effect of dynamical vegetation change. In these experiments with the suffix “fix”, model does not contain carbon/dynamic modules. Vegetation distribution is fixed to the equilibrium state of experiment CTRL. After 150 years AGCM run, monthly temperature, precipitation and cloud cover ratio in the last 20 years is used to drive offline LPJ-DGVM experiment in order to compare the equilibrium terrestrial carbon storage. This offline experiments are done for CTRL, 4xCO₂ and 4xCO₂fix.

3. Results

The equilibrium vegetation distribution in CTRL, 2xCO₂ and 4xCO₂ are shown in Fig. 2a, 2b, and 2c, respectively. Major differences of the vegetation distribution due to increased atmospheric CO₂ are seen as follows: (i) Vanishing of tundra in northern high latitudes, (ii) Northward shift of boreal forests in northern America and Eurasia, (iii) Poleward expansion of temperate and tropical forests, and (iv) Greening of arid/semi-arid areas in Central Eurasia, Australia, Sahel, Arabia and Near East region.

Fig. 2c, 2d, and 2e show the distribution of the net primary productivity (NPP) in CTRL, the change in NPP in 2xCO₂ and 4xCO₂. It is shown that NPP increases generally all over the land due to increased atmospheric CO₂. Fig. 3a and 3c shows the annually averaged surface temperature change in 2xCO₂ (2xCO₂ -CTRL) and 4xCO₂ (4xCO₂ -CTRL), respectively. It is shown that the temperature change over land ranges from +8K in the tropical region to +20K in boreal region due to quadrupled CO₂ and +4K to +10K due to doubled CO₂. The pattern of the change reflects the polar amplification which is a common feature in previous studies (IPCC TAR 2001). The temperature change due to quadrupled CO₂ generally resembles the change due to doubled CO₂ in pattern. The ratio (4xCO₂ -CTRL)/(2xCO₂ -CTRL) falls into the range of 1.6 to 2.4 in the majority of the land and sea surface (not shown). Hence, it is also shown that the magnitude of change due to the doubled/quadrupled atmospheric CO₂ is generally linear.
Fig. 2. Equilibrium vegetation distribution in experiment (a) CTRL, (b) 2xCO2 and (c) 4xCO2, (d) Equilibrium net primary productivity (gC/m**2) in CTRL, (e) difference between 2xCO2 and CTRL and (f) difference between 4xCO2 and CTRL.

The contribution of dynamic vegetation is shown in Fig. 3b (2xCO2 - CTRL) - (2xCO2fix - CTRLfix) and 3d (4xCO2 - CTRL) - (4xCO2fix - CTRLfix), respectively. It is shown that the dynamic vegetation contributes positive and enhances warming. This amplification of warming shows generally similar pattern to the total temperature change for both doubled and quadrupled case. Globally averaged surface air temperature warming increased from +4.3K to +4.7K (doubled CO2), and +8.7K to +9.6K (quadrupled CO2) by the inclusion of dynamic vegetation. This amplification is about +20% averaged all over the land in both doubled and quadrupled experiments.
Fig. 3. (a) Annually averaged warming (K) in doubled CO2 (2xCO2-CTRL) and as same as (c) in quadrupled CO2 (4xCO2-CTRL). (b) Difference of warming (K) between dynamic vegetation case and fixed vegetation case in doubled CO2 {(2xCO2-CTRL)-(2xCO2fix-CTRLfix)} and as same as (d) in quadrupled CO2 {(4xCO2-CTRL)-(4xCO2fix-CTRLfix)}.

Fig. 4. Albedo difference (%) due to the inclusion of dynamic vegetation in experiment 4xCO2{(4xCO2-CTRL)-(4xCO2fix-CTRLfix)}.

Fig. 4 shows the contribution of dynamic vegetation to albedo change in experiment 4xCO2. This difference is due to greening and vegetation shift shown in Fig. 2c. Albedo change and warming over land triggers ice-albedo feedback in Arctic Sea (not shown) which
reduces surface albedo. The pattern of change is similar to the amplification of warming (Fig.3d). We also compared change of equilibrium terrestrial carbon storage from CTRL in $4xCO_2$ and $4xCO_2fix$. Globally averaged vegetation and soil carbon storage difference from CTRL are reduced from +575 GtC to +497 GtC and from -358 GtC to -608 GtC due to the inclusion of dynamic vegetation respectively. The total difference of terrestrial carbon storage (-328 GtC) corresponds to the emission of 115 ppm carbon dioxide from land to atmosphere, which corresponds to 18% increase of carbon dioxide concentration difference in $4xCO_2$ and CTRL. The difference of soil carbon (Fig. 5) is major part of the change of terrestrial carbon storage. It is shown that the difference due to dynamic vegetation is seen in northern middle and high latitudes. This difference reflects the acceleration of soil carbon decomposition due to the amplified warming.

![Soil carbon difference (kgC/m**2) between 4xCO2 and 4xCO2fix.](image)

Fig. 5. Equilibrium soil carbon difference (kgC/m**2) between 4xCO2 and 4xCO2fix.

4. Summary

In order to investigate Influence of dynamic vegetation change on climate change due to increase of CO$_2$, we have coupled LPJ-DGVM to MIROC-AGCM. We examined preindustrial and quadrupled CO$_2$ equilibrium experiments using LPJ-MIROC, and preindustrial and quadrupled CO$_2$ equilibrium experiments using usual MIROC with fixed vegetation. The result shows potential vegetation change in doubled and quadrupled atmospheric CO$_2$ condition. The result also shows amplification of global warming caused by this vegetation change compared to fixed vegetation experiments. The warming is amplified by 10% globally due to the inclusion of dynamic vegetation, and amplified 20% over the land. It is also indicated that terrestrial carbon storage in high atmospheric CO$_2$ is significantly reduced by the inclusion of dynamic vegetation. Taking into account this reduction of terrestrial carbon storage, atmospheric CO$_2$ concentration increases by 18% and leads to further warming. Feedbacks shown in the result of the present study are able to explain discrepancy among global carbon cycle models partly.
References


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Solar Signals of 11-Year Cycle in Temperature and Ozone in the Middle Atmosphere Simulated with a Chemistry-Climate Model of Meteorological Research Institute

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Abstract
The 11-year solar signal in the middle atmosphere is investigated with a chemistry-climate model (CCM) of Meteorological Research Institute (MRI). The CCM employs spectral T42 truncation and 68 layers, and contains major stratospheric chemical species including HOx, NOx, ClOx, and BrOx. The CCM is integrated for about 25 years from 1979 to 2004 under the CCMVal REF1 scenario, in which all the forcings for the CCM are specified from observations. The forcings include sea surface temperature, sea ice, abundances of greenhouse gases and halogens at the surface, solar irradiance, and volcanic aerosols. Multiple regression analysis was made for the CCM results to select relevant solar signals. It is found that the CCM qualitatively reproduced the observed features in temperature and ozone solar signals in low latitude stratosphere.

Keywords: 11-year solar cycle, Multiple regression, Temperature solar signal, Ozone solar signal, Chemistry-climate model

1. Introduction

The solar radiation is the primary energy source for the dynamical structure and chemical composition of the earth’s atmosphere. Its variations hence give rise to variations not only in the upper atmosphere but also in lower and near-surface atmosphere. Above all, the 11-year solar cycle, one of the most prominent components, accompanies substantial change (several per cent or more) at the ultra-violet wavelengths, though visible and infrared irradiance changes are very small with the total solar irradiance (TSI) being about 0.1%.

Analyses of upper atmosphere data based on satellite observations have proved that the 11-year solar cycle changes the upper stratospheric temperature over the equator by 0.5 to 1.8 K (e.g., Labitzke et al., 2002; Scaife et al., 2000; Crooks and Gray, 2005). Ozone is also analyzed to vary with the 11-year solar cycle in the midlatitude upper stratosphere by, at most, 4.5% (Hood, 1997; Shindell et al., 1999). However, there are large uncertainties in these evaluations of solar signals because the entire length of satellite observations barely covers only three solar cycles and major volcanic eruptions (El Chichón in 1983 and Mount Pinatubo in 1991) overlapped with the solar maximum phases.

Numerical simulations of the 11-year solar cycle have been made with 2-dimentional models (e.g., Fleming et al., 1995; Lee and Smith, 2003) and 3-dimentional models (Tournape, 2003; Egorova et al., 2004, 2005), which includes chemistry. Note that 3-dimentional models with chemistry and radiative feedback of radiatively active gases are referred to chemistry-climate models (CCMs).

However, there are virtually large discrepancies between observations and such models. One reason for this is that the ozone radiative feedback is not well represented in the models.
due probably to the deficit in the wave-mean flow interaction treatment, which deficit comes from systematic errors in basic thermal structure.

The purpose of this study was to investigate to what extent MRI-CCM reproduces the 11-year solar signal in temperature and ozone through dynamics and chemistry by imposing realistic forcings for the past 25-year from 1979 to 2004. It is found that the CCM qualitatively reproduced the observed features in temperature and ozone solar signals in low latitude stratosphere.

2. Model and Experiments

The chemical module of MRI-CCM (Shibata et al., 2005) used in this study is an updated version, in which the transport scheme is improved and fluorine chemistry is incorporated. The vertical transport of a cubic expression for column abundances from the top is replaced to the piecewise rational method (RPM) (Xiao and Peng, 2004) and the horizontal interpolation of chemical species abundances is changed from a cubic and liner pair to a quintic and cubic pair for Lagrangian interpolation. The dynamical module is a T42L68 general circulation model, in which the non-orographic gravity wave drag scheme by Hines (1997) is used with an enhanced source in the tropics, and the vertical spacing of layers is 500 m between 100 to 10 hPa (Shibata and Deushi, 2005a). The chemical module (Shibata et al., 2005) contains major stratospheric species, e.g., 36 long-lived and 15 short-lived species with 80 gas-phase reactions, 35 photodissociations, 6 and 3 heterogeneous reactions on polar stratospheric clouds and sulfate aerosols.

Single time integration of the CCM, starting from November 1979, is made to the end of 2004 by imposing realistic forcings. Since the details of the forcings are described in Eyring et al. (2006) and references therein, only brief explanation is given here. Sea surface temperature (SST) and sea-ice is taken from Met Office Hadley Centre’s data archive: HADISST1 (Rayner et al., 2003), and abundances of greenhouse gases and halogens are specified at the surface based on IPCC (2001) and WMO (2003), respectively. Volcanic aerosols are derived from updated data of Sato et al. (1993) for extinction and effective radius, and from updated data similar to that used by Jackman et al. (1996) for surface area. Solar irradiance of 1 nm spectral interval is used to make a lookup-table for photodissociation as in Sekiyama et al. (2006) and to create radiative fluxes in sub-bands in radiation scheme. All these forcings are monthly-mean data, from which daily data is temporally interpolated. Note that no particular treatment for the QBO was made, because the T42L68 version reproduces the QBO in zonal wind (Shibata and Deushi, 2005a) and in ozone (Shibata and Deushi, 2005b) as well.

The multiple regression analysis (e.g., Randel and Cobb, 1994; Bodeker et al., 2001) is applied for the anomaly time series of model and observed data such as,

\[ O_3(t) = A(NA=3) + B(NB=3) \times t + C(NC=3) \times \text{QBO}(t) \text{ (at P=50hPa)} + D(ND=3) \times \text{QBO}(t) \text{ (at P=20hPa)} + E(NE=3) \times \text{Pinatubo}(t) + F(NF=3) \times \text{El Chichón}(t) + G(NG=3) \times \text{ENSO}(t) \]
+ H(NH=0) × Solar Flux (t)  
+ Residual (t)

where QBO represents zonal wind, the global mean extinctions of volcanic aerosols at 0.55 micron meters of Mount Pinatubo and El Chichón, ENSO the southern oscillation index, and Solar Flux the 10.7-cm radio flux (F10.7) in solar unit \((10^{-22} \text{ Wm}^{-2}\text{Hz}^{-1})\). The coefficients A through G are expanded to explain seasonality as

\[
A = A_0 + A_1 \cos(\omega t) + A_2 \sin(\omega t) + A_3 \cos(2\omega t) + A_4 \sin(2\omega t) + A_5 \cos(3\omega t) + A_6 \sin(3\omega t)
\]

\[\omega = 2\pi / (12 \text{ months}).\]

Note that the coefficient of solar flux does not include seasonality, indicating its coefficient being annual-mean.

3. Temperature Solar Signal

Fig. 1 displays the annual-mean temperature solar signal (K per 100 units of F10.7) from 100 to 0.1 hPa equatorward of 60 degrees latitudes for the CCM and ERA-40 reanalyses (Uppala et al., 2005). The period for the multiple regression is 1980-2004 for the CCM and 1980-2001 for ERA-40. In the simulation there appear two warming peaks in low latitudes: the maximum around 0.5 hPa in the mesosphere and the second one around 70 hPa in the low stratosphere. The mesospheric and upper stratospheric strong warming area has larger latitudinal extent but shallower vertical depth than the lower stratospheric warming. Note that there is weak cooling area around 7 hPa between the two warming areas, though statistically insignificant. In the ERA-40 reanalyses, the temperature solar signal is larger than that of the simulation, though there is also similar three-cell pattern (positive-negative-positive) in low latitudes. Above all, the maximum warming in the upper stratosphere is much underestimated.

![Fig. 1. Latitude-pressure cross-section of solar regression coefficients of temperature (K/F10.7*100) derived from (left) the model and (right) from ERA-40. Contour interval is 0.1 for the model and 0.2 for ERA-40. Shading denotes the area statistically significant at 95% level.](image-url)
and its position is much higher in the simulation. On the other hand, the position of this upper stratospheric warming is well simulated with MRI-GCM in the time-slice run of solar maximum and solar minimum under prescribed ozone change (Shibata and Kodera, 2005), indicating that the ozone feedback effect is responsible for this discrepancy between the simulation and reanalyses. In should be here stressed that the temperature solar signal is dependent on the detail procedure of the multiple regression analysis, because different multiple regression results in different solar signal even when the same ERA-40 datasets are used (Crooks and Gray, 2005).

4. Ozone Solar Signal

Fig. 2 exhibits the annual-mean ozone solar signal (% per 100 units of F10.7) from 100 to 0.1 hPa equatorward of 60 degrees latitudes for the CCM and Solar Backscattered Ultraviolet (SBUV) version 8 (updated from Miller et al. [2002]). SBUV period used in this study is from 1980 to 2001. The simulation shows four-cell pattern, negative (-0.8%) about 0.5 hPa, positive (1.2%) about 3 hPa, negative (-0.6%) around 30 hPa, and positive (1.6%) around 70 hPa in the low latitude stratosphere, though the last one is statistically insignificant. The ozone decrease in the mesosphere is due to the increase of HOx arising from enhanced irradiance at Lyman-a line.

The SBUV ozone data has similar but higher-altitudes three-cell structure in low latitudes, while there is no hint of the mesospheric negative cell. In addition, the ozone solar signal of SBUV is much larger than that of the simulation and there are positive cells around 10 hPa in both midlatitudes (35S and 35N). This structure is consistent with the ones referred to “doubled-lobed” (McCormack and Hood, 1996) or “dipole” (Lee and Smith, 2003). The midlatitudes positive cells are also seen in the simulation, but their altitudes are higher and statistically insignificant.

![Fig. 2. Latitude-pressure cross-section of solar regression coefficients of ozone (%/F10.7*100) derived from (left) the model and (right) from SBUV. Contour interval is 0.2 for the model and 0.3 for SBUV. Shading denotes the area statistically significant at 95% level.](image-url)
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The Influence of Volcanic Activity and Changes in Solar Irradiance on Surface Air Temperatures in the Early Twentieth Century

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Abstract
Causes of the global surface air temperature warming in the early half of the 20th century are examined using a climate model and an optimal detection/attribution methodology. While the anthropogenic response seems to be underestimated in our model, our previous study detected the influence due to natural external forcing, including the combined effects of solar irradiance changes and the recovery from large volcanic activity. We further partition the responses between these two natural external factors, detecting both the solar and the volcanic signal in the observed early warming. A diagnosis of the sensitivity to solar forcing and a volcanic super-eruption simulation suggest that our model possesses larger climate sensitivities to solar forcing and longer relaxation times to volcanic forcing than HadCM3, enabling us to detect both the solar and volcanic forcing responses.

Keywords: Climate change, Global warming, Detection and attribution

1. Introduction
By comparing coupled atmosphere-ocean general circulation model (CGCM) simulations to observations, climate modelling groups have made efforts to understand the reason of the observed global surface air temperature (SAT) warming in the early half period of the 20th century. However, the factors causing the observed early century warming have remained a topic of debate. Stott et al. (2003, hereafter S03) indicated that, using the simulations with HadCM3 (Gordon et al., 2000), the early warming is attributable to a combination of increased well-mixed greenhouse gases (GHGS) and solar irradiance changes. Hegerl et al. (2003) detected both the GHGS and volcanic signals. Delworth and Knutson (2000) mentioned the importance of the GHGS influence and internal variability. On the other hand, our previous study (Nozawa et al., 2005, hereafter N05) suggested that, using MIROC3.2 (K-1 model developers, 2004), natural factors (consisting of solar irradiance change and volcanic influence) are important for the early warming. It would be noted that, in N05, the uncertainty range of the GHGS signal is not well constrained, although its best-estimate amplitude is similar to that of the observations.

It is clear that the exact cause of the early century warming is still uncertain and requires further investigations. Understanding the cause of past climate change is imperative, since it allows us to check the ability of CGCMs to model the climate system, and to provide evidence of the reliability of our projections of the future climate change with those CGCMs. In this study, we will promote the work of N05 by performing additional simulations of MIROC3.2 to investigate the relative importance of solar and volcanic forcings on the early century warming. We’ll also discuss the reason why the anthropogenic contribution is not
well constrained in our model. Furthermore, we will explore why different CGCMs respond differently to solar and volcanic, and why these different responses lead to different estimates of the relative contributions of different factors causing the early century warming.

2. Simulations, Observations, and Methods

We consider four ensembles of 20th century climate simulations performed by MIROC3.2. The first ensemble was forced with changes in solar irradiance (Lean et al., 1995) alone (hereafter SOLR); the second with changes in stratospheric aerosols due to volcanic activity (Sato et al., 1993) alone (VLCN); the third with the influence of both solar and volcanic factors (NTRL); and the last with changes in anthropogenic factors (ANTH; see N05 for the detail). Each of the ensembles consists of four ensemble members with initial conditions taken at 100-yr intervals from a stable 1300-yr pre-industrial control simulation.

The model simulated SAT changes are compared with the dataset of Jones and Moberg (2003) which contains monthly mean SAT from station observations amalgamated onto a 5° lat. × 5° lon. grid. Our simulated SAT data are bi-linearly interpolated in latitude and longitude from its original T42 grid to the observational grid, discarding at time and locations where the observations were not available.

To determine the relative importance of the external factors on the early 20th century warming, the optimal detection/attribution approach (Allen and Tett, 1999) is applied to the model simulations and the observations for the 1900–1949 period. Using this method, we express the spatio-temporal pattern of the observed SAT changes as a linear sum of a couple of scaled SAT changes forced by external factors plus internal climate noise. Scaling factors are computed to provide the best fit to the observations. Taking into account the climate noise included in the finite ensemble mean of model simulated responses, we apply a total least squares (TLS) regression (Allen and Stott, 2003).

For the detection/attribution analysis, 10 pentad (5-yr) mean anomalies during the 1900–1949 period are used rather than the decadal (10-yr) mean diagnostic used in N05, since typical recovering time from volcanic cooling is considered to be shorter than a decade. We require ≥ 5 pentads at each grid point; otherwise the pentad averages are omitted from the analysis. These anomalies are spatially truncated onto T4 spherical harmonics.

For signal-to-noise optimization, the observed and simulated SAT anomalies are projected onto the leading 20 eigenvectors of intra-ensemble variability (the variability between the four simulations, after subtracting the mean signal), rather than 10 eigenvectors used in N05, since the pentad mean data analyzed in the present study have twice the temporal resolution of the decadal mean data investigated in N05. The results given below are insensitive to the exact level of truncation. The uncertainties on the signal amplitudes are estimated from a 1300-yr control simulation. The residual test of Allen and Tett (1999) indicates that the residual in the regression is not inconsistent with control variability. Furthermore an $F$-test indicates no significant inconsistency between the variability of the global pentad mean SAT in the control simulation and that in the detrended observations at the 10% level.
Fig. 1. Temporal changes of global annual mean surface air temperature. Black thick lines are anomalies from the 1881-1910 mean for the observations, red thick lines indicate the ensemble mean of the ANTH, NTRL, SOLR, and VLCN simulations. Maximum and minimum ranges from the individual simulations are shaded in pink.

3. Detection and Attribution

Fig. 1 shows temporal variations of the global annual mean SAT for the observations and the model simulations. The observed warming over the early half of the 20th century is well simulated by NTRL, whereas ANTH show little warming. Both SOLR and VLCN produce a global warming during the same period, although the magnitude of the warming is less than that for the observations. In the SOLR and VLCN simulations, these warming trends can be explained by the increase of solar irradiance over the early 20th century (Lea et al., 1995), and a recovery from the cooling due to the heavy volcanism in the 1900s and the 1910s (Sato et al., 1993), respectively.

Results of the 2-way (SOLR and VLCN) and 3-way (ANTH, SOLR, and VLCN) optimal detection/attribution analyses are given in Figs. 2 and 3, respectively. Fig. 2a depicts the scaling factors with their uncertainty ranges for SOLR and VLCN. The uncertainty ranges of both SOLR and VLCN signals are positive, and include unity, indicating that these simulated signals are detectable in the observed SAT changes, and their signal amplitudes are consistent with that in the observations.

Best-estimate trends of global mean SAT are shown in Fig. 2b. The observed trend (= 0.9 K/century) is well reproduced by a combination of the SOLR and VLCN signals. Each of the scaled trends for SOLR and VLCN is about a half of the observed trend (their best estimates are 0.34 K/century and 0.46 K/century, respectively). These results suggest that the global warming over the early half of the 20th century is attributable to both effects of the solar irradiance change and the large volcanic activity, at least in MIROC3.2.
We also perform the ANTH, SOLR, and VLCN 3-way detection/attribution analysis (Fig. 3). The SOLR and VLCN signals show no changes from the 2-way analysis (Fig. 2), while the uncertainty range of the ANTH signal is poorly constrained as discussed by N05. Therefore the effects of the anthropogenic forcing are unclear, while the solar and volcanic responses are robust. Squared signal-to-noise ratio, \((SNR)^2\) defined by Tett et al. (2002), of the ANTH response is 1.4, which is not significant at the one-sided 10% level of the \(F\)-test. This poor \((SNR)^2\) prevents us from achieving good estimates of the uncertainty range of the ANTH scaling factor. On the other hand, \((SNR)^2\) for SOLR and VLCN (2.0 and 2.1, respectively) are significant at the 10% level.

It should be noted that, although the ANTH signal is not detected during the early 20th century with MIROC3.2, it is detected in a full-century or 1950–99 analysis. For example, a detection/attribution analysis of the decadal mean ANTH and NTRL signals for 1900–99 shows that their scaling factors are 1.81 < its 5–95% uncertainty range is 1.38 · 2.40> and 1.03 <0.28 · 2.01>, respectively. However, it is also suggested that MIROC3.2 slightly underestimates the ANTH response, while the NTRL one is consistent with that in the observations. This underestimation of the ANTH response is related with its less \((SNR)^2\), making the estimate of the ANTH contribution unclear. Although it is important to investigate the reason why the ANTH response in MIROC3.2 is small, it is beyond the focus of the present study.

There is uncertainty in the solar and volcanic forcing in the early part of the 20th century, and the solar forcing seems to be less certain than volcanic forcing (e.g., Foukal et al., 2004; Ammann et al., 2003). To test the sensitivity of our results, it is worthwhile to analyze additional ensembles forced with other reconstructions of solar and volcanic forcing.

**Fig. 2.** Best-estimates (a) of scaling factors and (b) linear trends (K/century) with 5-95% uncertainty ranges for the SOLR and VLCN in a 2-way analysis. The orange error bar with an asterisk and purple error bar with a triangle indicate SOLR and VLCN, respectively. Also shown in (b) are the total reconstructed trend from the regression model (yellow error bar with a plus) and the observed trend (black error bar with a square).

**Fig. 3.** Same as Fig. 2., except for the ANTH, SOLR and VLCN 3-way analysis. The green error bar with a diamond indicates the estimated ANTH signal.
4. Comparison with the HadCM3 Model

The reconstructed solar response trend of MIROC3.2, 0.34 $<0.04 \cdot 0.68>$ K/century, is consistent with the results of S03, who found a trend due to solar of 0.29 $<0.15 \cdot 0.46>$ K/century from simulations of HadCM3 forced with solar irradiance changes according to Lean et al. (1995). However, in HadCM3 the scaling factor for the solar signal is 2.64 (obtained from 1900–99 analysis), compared to 1.19 in MIROC3.2, hence suggesting that the non-scaled response in MIROC3.2 to solar forcing is larger than that in HadCM3. Since the response to solar forcing in MIROC3.2 is consistent with the observed response, it shows that we can obtain the suitable magnitude of response to solar forcing without recourse to an additional physical process which may act to amplify the amplitude of the solar response, such as a solar-ozone feedback process (Haigh, 1996). However, it does not preclude that these additional influences are important and may have a discernible impact on SATs.

For the sake of future projections (Stott and Kettleborough, 2002), it is important to understand why these two models respond differently to the solar forcing. The diagnosing method of climate feedback parameter ($\alpha$) derived by Gregory et al. (2004) enables us to compare sensitivities ($S$) of MIROC3.2 to solar forcing and that of HadCM3. Note that smaller $\alpha$ indicates greater $S$, since $S$ is proportional to $\alpha^{-1}$. The $\alpha$ of MIROC3.2 to solar irradiance increases is 1.07, which is less than that of HadCM3 (2.0; see Gregory et al., 2004), demonstrating that MIROC3.2 has larger sensitivities to solar forcing than HadCM3. Larger sensitivities would cause the larger responses to the historical solar irradiance changes, making it easier to detect the solar signal. It would be worth stressing that MIROC3.2's sensitivity to solar forcing ($\alpha = 1.07$) is consistent with its sensitivity to GHGS forcing ($\alpha = 1.13$), whereas the HadCM3 sensitivities are inconsistent ($\alpha = 2.0$ for solar, and $\alpha = 1.26$ for GHGS), so this may highlight a puzzle for HadCM3.

There is also a large difference in the volcanic responses between the present study and S03, i.e., the volcanic signal is detectable in the present study, while that was not detected in S03. Forster and Collins (2004) and Yokohata et al. (2005) demonstrated that both HadCM3 and MIROC3.2 reproduce the SAT response to the Mt. Pinatubo eruption in June 1991 in good agreement with the observations. To compare the response in both the CGCMs to a volcanic activity under better signal to noise ratios, we performed the volcanic super-eruption experiment (Jones et al., 2005), which simulates the response to the aerosol optical depth changes following a volcanic activity which is 100 times larger than that of the Mt. Pinatubo eruption. MIROC3.2 has longer relaxation time to the volcanic super-eruption (recovering to a half amplitude of peak cooling by $\approx 7$ yrs) than HadCM3 ($\approx 5$ yrs; see Jones et al., 2005), although the peak cooling amplitude of MIROC3.2 (9.7 K) is slightly less than that of HadCM3 (10.7 K) (not shown). This difference of the relaxation time to the volcanic eruption may cause the different estimates of the volcanic contribution to the early century warming: with the model possessing longer relaxation time, the model volcanic signal tends to be less smoothed out by the temporal low-pass filter (pentad mean) applied in the detection/attribution analysis, likely resulting in that the volcanic signal is easier to be detected. However, it should be noted that a volcanic super-eruption might, because of non-linear changes in ocean circulation, not be representative of the response time for more regular eruptions.
5. Summary

We investigated the relative roles of solar and volcanic contributions on the global warming over the early half of the 20th century. By applying the optimal detection/attribution methodology to the observations and the MIROC3.2 simulations, it was suggested that both the solar and volcanic influences are detectable in the observed early warming. On the other hand, the uncertainty in the anthropogenic contribution is not successfully constrained, since its signal is probably underestimated.

Climate responses to solar and volcanic forcing in MIROC3.2 were compared with HadCM3. MIROC3.2 has larger climate sensitivities to the solar forcing than HadCM3 by a factor of 2. Unlike HadCM3, the solar sensitivity of MIROC3.2 is consistent with its greenhouse gases sensitivity, and therefore, MIROC3.2 seems to be telling a simpler story than HadCM3. Furthermore, MIROC3.2 has longer relaxation times to volcanic cooling than HadCM3. These differences would affect the detection/attribution of the past climate change, and also estimates of the uncertainty of the future climate projection. Since it is still unclear why there are large uncertainties in climate responses to solar and volcanic forcings, it is necessary to compare mechanisms determining climate responses to external forcing factors among various CGCMs.

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**Original Papers and Reviews:**


**Conference Reports:**


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2. Atmospheric and Oceanic Environment Modeling
Numerical Simulation of Thermal and Airflow Field around Regularly Arrayed Buildings

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Abstract
Urban canopy models can take into account the properties of urban area more precisely than roughness models and require smaller computational costs than bluff body models which resolve buildings when computing the flow field. So, applications of urban canopy models to urban climate simulations for the evaluation of heat island countermeasures are expected to be useful. However, the accuracy of urban canopy models, especially for temperature, has not been verified. Therefore, in this study, we performed numerical simulations of the flow around regularly arrayed buildings. We obtained data on horizontally space-averaged wind velocity and air temperature, which is necessary for the development and verification of urban canopy models. These numerical results were compared to measurements taken in a thermally stratified wind tunnel.

Keywords: Thermal and airflow field, k-ε model, Regularly arrayed buildings

1. Introduction

Recently, countermeasures against urban heat island phenomena have become political issues and investigation of the effects of these countermeasures is very important. From the point of view of computational cost and ability to take into account urban properties, urban canopy models (Hiraoka et al., 1990, Maruyama, 1994, Vu et al., 2002) are considered to be one of the most practical simulation tools for the investigation of these countermeasures. However, the accuracy of urban canopy models has not been satisfactorily verified especially for their ability to simulate temperature. Furthermore, there is almost no database for the verification of urban canopy models especially for temperature.

In this study, to prepare for verification of urban canopy models, we obtained the data of horizontally space-averaged wind velocity and air temperature, by performing numerical simulations of airflow and air temperature around regularly arrayed buildings. Moreover, we investigated the accuracy of the obtained data by comparing them with the measured data from thermally stratified wind tunnel experiment.

2. Numerical Method

The governing equations are shown in Table 1. To solve the governing equations, the present study employs the SMAC method, and discretizations are performed by the finite difference method on a Cartesian staggered grid system. The convection term is discretized by
a first-order upwind differencing scheme and diffusion term is discretized by a second-order central differencing scheme. Time advancing is performed by the implicit Euler method. The Poisson equation for pressure is solved using the incomplete Cholesky conjugate gradient (ICCG) method.

Table 1 Governing Equations

<table>
<thead>
<tr>
<th>Equation Type</th>
<th>Equation</th>
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</thead>
<tbody>
<tr>
<td>(continuity equation)</td>
<td>( \frac{\partial u_i}{\partial x_i} = 0 )</td>
</tr>
<tr>
<td>(momentum equation)</td>
<td>( \frac{\partial u_i}{\partial t} + \frac{\partial u_i u_j}{\partial x_j} = -\frac{1}{\rho} \frac{\partial P}{\partial x_i} + \frac{\partial}{\partial x_j} \left[ \nu_t \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \right] )</td>
</tr>
<tr>
<td>(energy equation)</td>
<td>( \frac{\partial T}{\partial t} + \frac{\partial u_i u_j}{\partial x_j} = \frac{\partial}{\partial x_j} \left[ \frac{\nu_t}{\sigma_e} \frac{\partial T}{\partial x_j} \right] )</td>
</tr>
<tr>
<td>(transport equation for ( k ))</td>
<td>( \frac{\partial k}{\partial t} + \frac{\partial u_i k}{\partial x_j} = \frac{\partial}{\partial x_j} \left[ \frac{\nu_t}{\sigma_k} \frac{\partial k}{\partial x_j} \right] + P_k + G_k - \varepsilon )</td>
</tr>
<tr>
<td>(transport equation for ( \varepsilon ))</td>
<td>( \frac{\partial \varepsilon}{\partial t} + \frac{\partial u_i \varepsilon}{\partial x_j} = \frac{\partial}{\partial x_j} \left[ \frac{\nu_t}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial x_j} \right] + \frac{k}{\varepsilon} \left( C_{1k} \frac{\partial k}{\partial x_j} - C_{2k} \varepsilon + C_{3k} \max[0, G_k] \right) )</td>
</tr>
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Here,

\( v_t = C_\mu \frac{k^2}{\varepsilon}, \quad P_k = v_t \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \frac{\partial u_i}{\partial x_j}, \quad G_k = \frac{v_t \rho_i}{Pr} \frac{\partial T}{\partial x_i} \)

\( u_i \): wind velocity in \( x_i \) direction, \( t \): time, \( \rho \): air density, \( P \): pressure function, \( T \): air temperature, 
\( k \): turbulent energy, \( \varepsilon \): viscosity dissipation rate, \( \nu_t \): eddy viscosity coefficient, \( Pr \): turbulent Prandtl number

\( C_\mu = 0.09, \sigma_k = 1.00, \sigma_\varepsilon = 1.30, \ C_{1k} = 1.44, \ C_{2k} = 1.92 \)

3. Computational Methodology

As shown in Fig.1, the computational domain is 1.7 m in the streamwise direction (X), 1 m in the lateral direction (Y) and 0.637 m in the vertical direction (Z). This domain is a part of the wind tunnel owned by the Building Research Institute of Japan, which has a working test section of length 10 m, width 1 m and height 1 m. Computational conditions are imposed according to the wind tunnel experiment (Komatsu et al., 2005), in which wind and air temperature distributions around an array of 3-D buildings were measured. This array consists of 7 rows of 10 sharp-edged wooden blocks, each 60 mm long and wide, and 40 mm high. The streamwise and spanwise spacing between blocks is 50 mm, giving a gross building ratio of 30%.

At the inlet boundary, the measured profiles of streamwise mean velocity, turbulence kinetic energy and air temperature are used. The maximum streamwise mean velocity is 1 m/s.
and the profile of the velocity accords well with a 0.25 power law profile which is a typical profile over cities. The air temperature is uniformly 25°C. The dissipation rate is estimated by assuming an equilibrium turbulent flow in which the rates of production and destruction of turbulence are in near balance. At the outlet boundary, the gradients of all variables are assumed to be zero in the flow direction. At the upper boundary, free-slip conditions are imposed for all variables. At all the solid boundaries (bottom boundary, lateral boundaries, block walls), standard wall functions are applied for the mean velocities and turbulence quantities. The temperature of the floor is 60°C except the 25°C area with 0.1 m length near inlet boundary, and the convective heat transfer boundary condition is implemented using a heat transfer coefficient of $\alpha_c = 11.6$ [W/m²K]. The lateral walls of the wind tunnel are assumed adiabatic and Neumann condition is implemented for air temperature. In order to consider the effects of thermal radiation and heat conduction from the floor on the temperature of block walls, following 5 cases are computed.

Case1: block wall temperature is 25°C, convective heat transfer boundary condition
Case2: block wall temperature is 30°C, convective heat transfer boundary condition
Case3: block wall is adiabatic, Neumann condition
Case4: block wall temperature is measured value, convective heat transfer boundary condition
Case5: block wall temperature is measured value, convective heat transfer boundary condition
Case4 and Case5 use the temperature measured at 3 to 4 points on each wall in conjunction with extrapolated temperatures. The heat transfer coefficient of Case1, Case2 and Case4 is $\alpha_c = 11.6$ [W/m²K]. As for Case5, in order to take into account the effect of local distribution of convective heat transfer coefficient, following formula proposed by Jürges (McAdams, 1954) is applied.

$$\alpha_c = 5.8 + 4.0 \times V_l [W/m^2K]$$ (1)

Here, $V_l$ is local wind velocity near the surface of blocks’ walls.

The grid spacing in the X and Y direction is 5 mm. The grid spacing in the vertical direction is stretched upward, and is 1 mm between the first two levels. The total number of grid points is $340 \times 200 \times 43 = 2,924,000$.

![Computational domain and boundary condition](image)

**Fig. 1. Computational domain and boundary condition.**
4. Results and Discussion

Fig. 2 shows the horizontal distribution of wind velocity at half the height of the blocks. Although the airflow around outermost rows of blocks is not symmetric because of the absence of adjacent rows of blocks and the effect of wind tunnel walls, the flow around the central blocks is symmetric. So, by spatially averaging physical quantities around the central blocks, we can obtain data for verification of urban canopy model. We performed horizontal space-averaging of wind velocity and air temperature in the area indicated by the red square in Fig. 2.

![Horizontal distribution of wind velocity and air temperature](image)

**Fig. 2. Horizontal distribution of wind velocity at half the height of blocks.**

Fig. 3 shows the vertical distributions of horizontally space-averaged wind velocity and air temperature. The vertical axis indicates the normalized height. The horizontal axis indicates normalized wind velocity (left) and normalized air temperature (right), respectively. Here, $H$ is the height of the block, $u_0$ is the wind velocity at 0.5 m above the floor, $\theta_f$ is floor temperature and $\theta_0$ is the air temperature at 0.5 m above the floor. As for the wind velocity, the computed results of all cases generally agree well with the measurements (WTT-data). On the other hand, the differences of computed air temperature among 5 cases are large and the conformance between some computed cases and measurements is not good as those for the streamwise wind velocity.

For further investigation of the flow field, vertical distributions of streamwise wind velocity and air temperature at point A (Aisle side) and point B (Behind building) depicted in Fig. 2 are displayed in Fig. 4 and Fig. 5. In Fig. 4, the wind velocity distributions of all cases at both point A and B agree well with measurements. Even the negative values ($u/u_0<0$) due to recirculation are predicted at point B. In contrast, at point B in Fig. 5, computed air temperature of all cases are much lower than measurements, implying poor simulation of thermal convection and diffusion in the region from the floor up to the block height. It is possible that this is because the wind velocity is so small behind blocks that the standard $k-\varepsilon$ model is not appropriate for the simulation in this region.
Fig. 3. Vertical distributions of horizontally space-averaged wind velocity (left) and air temperature (right).

a) A: Aisle side
b) B: Behind building

Fig. 4. Vertical distributions of local wind velocity.

a) A: Aisle side
b) B: Behind building

Fig. 5. Vertical distributions of local air temperature.
5. Summary

In this study, we performed numerical simulations of the airflow and air temperature around regularly arrayed blocks and obtained the data of horizontally space-averaged wind velocity and air temperature, which is necessary for the development and verification of urban canopy models. By comparing these data with wind tunnel measurement data, it was confirmed that the accuracy of wind velocity is sufficient to be used for the verification of urban canopy models. However, the accuracy of air temperature is not satisfactory and further improvement is needed for the simulation of flow field behind blocks.

In the future, in addition to the foregoing improvement of CFD code, we are planning to investigate the validity of an urban canopy model using the data obtained through this study.

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Cloud Droplet Collisions in Turbulent Mesoscale Convective Clouds

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Abstract

There is a growing consensus that collision growth rate of cloud droplets can be increased by turbulence. First, we developed a collision kernel model to predict the collision frequency in cloud turbulence. Then, we developed and performed numerical simulations of convective clouds to investigate the turbulence effects on cloud droplet collisions. Our preliminary predictions show that the particle collision growth is dramatically promoted by turbulence in orographic convective clouds and that the total amount of rainfall over a mountain is significantly increased.

Keywords: Cloud microphysics, Particle collisions, Two-phase turbulent flow, Large-eddy simulation

1. Introduction

Collision and coagulation processes of particles suspended in turbulence are often seen in both environmental and industrial flows. Typical examples are the growth of liquid droplets in turbulent clouds (Saffman and Turner, 1956; Pinsky and Khain, 1997) or in wet steam generators and spray atomization process. In these flows, the particle density is much larger than that of the ambient carrier fluid and the particle size widely distributes. When turbulence level corresponding to the averaged kinetic energy dissipation rate, $\varepsilon$, is large and the time scale for small-scale turbulence (the Kolmogorov time scale, $\tau_k$) is small, the particle inertial effects that are represented by the ratio of particle relaxation time $\tau_p$ to $\tau_k$ tend to be significant. Therefore inertial effects resulting from dynamic interactions of particles with turbulent fluid motion at finite Stokes numbers, $St \ (= \tau_p/\tau_k)$, must be considered as an important factor in the turbulent coagulation process.

The overall coagulation rate of finite-size particles in fluid turbulence is governed by three consecutive and interrelated processes: (i) geometric collision due to turbulent advection and gravitational settling, (ii) collision efficiency due to local particle-particle aerodynamic interactions and (iii) coagulation efficiency as determined by surface sticking characteristics. Here we focus on the geometric collision mentioned in (i), for which two inertia-induced mechanisms have been identified: (1) the relative velocities between two colliding particles due to turbulent fluctuation of the ambient carrier fluid (Saffman and Turner, 1956; Williams and Crane, 1983) and (2) the tendency of inertial particles to concentrate on certain areas in turbulent flow (Maxey 1987; Squires and Eaton, 1991; Sundaram and Collins, 1997). We refer to the first mechanism as “turbulent transport effect” and the second as “accumulation effect”. It has been demonstrated that both the inertia-induced effects can lead to one or two orders of magnitude increase in the average collision kernel (Sundaram and Collins, 1997;
Wang et al., 2000) and cause a much faster broadening of droplet size spectrum than previously believed.

Recent studies suggested that the collision frequency of cloud droplets can be remarkably increased by turbulence (Shaw 2003; Falkovich et al., 2002; Lynn et al., 2005). However, it is still controversial to conclude how significant the turbulence effect on droplet collisions is in real mesoscale clouds. In order to conclude this turbulence effect, it is needed to simulate the cloud collision growth with considering the droplet size explicitly. The spectral (bin) microphysics mesoscale model can potentially be used for this purpose. However, computer limitations have precluded the development of such mesoscale three-dimensional models with spectral microphysics.

Lynn et al. (2005) simulated a squall line developed over Florida using the three-dimensional fifth-generation Penn State-NCAR Mesoscale Model (MM5) based on spectral bin microphysics. They investigated the turbulence effects on collisions of cloud droplets and reported that the inclusion of the turbulence effect improves the timing and development of precipitation. Their simulation was the first attempt to investigate the turbulence effects in mesoscale clouds over real topography. However, their study has had some insufficient points. First, they used enhancement factors to describe the turbulence collision enhancement without accounting for possible changes in turbulent intensity depending on cloud evolution. Second, they used a horizontal numerical resolution of 3 km, which is not enough fine to capture the atmospheric turbulence. Third, they just compared the simulation results with observations of rainfall averaged over some observation sites. Due to the three points, turbulence effects on collisions could have been partly masked.

The purpose of this study is, therefore, to clarify the turbulence effects on cloud droplet collisions in mesoscale convective clouds. First, we developed a collision kernel model to predict the collision frequency in cloud turbulence. Then, we developed a new bin microphysics model that can explicitly predict the change of droplet spectrum broadening due to the turbulent collisions. We conducted the mesoscale simulations with sufficiently high resolution to capture the atmospheric turbulence.

2. Turbulence Effects on the Particle Collision Growth

2.1 Stochastic Collision Equation

The change rate of particle number density, \( n_i(r, x, t) \), by the stochastic coalescence process is expressed by

\[
\left( \frac{\partial n_i(r, x, t)}{\partial t} \right)_{\text{col}} = \frac{1}{2} \int_0^r K_c(r, r') n_i(r') n_i(r') dr' - \int_0^r K_c(r, r') n_i(r) n_i(r') dr',
\]

where \( r'' = (r^2 - r^2)^{1/3} \) and \( K_c(r_1, r_2) \) is the collision kernel describing the rate at which a particle of radius \( r_1 \) is collected by a particle of radius \( r_2 \). A model for the collision kernel is needed to predict the particle collision growth.

2.2 Collision Kernel Models

The conventional collision kernel model is the hydrodynamic kernel model, which describes the collision due to the settling velocity difference between two different size particles;

\[
\left\langle K_{c, \text{hyd}, 12} \right\rangle = \pi R_{12}^2 |V_{\infty, 1} - V_{\infty, 2}|
\]

where \( < > \) denotes an ensemble average, \( R_{12} = r_1 + r_2 \) is the collision radius and \( V_{\infty, j} \) is the
settling velocity of particles with radius \( r \). This hydrodynamic kernel cannot describe the collisions due to turbulence since it contains no flow parameter.

The average collision kernel, which involves the turbulence effects, can be written in the following form (Sundaram and Collins, 1997; Zhou et al., 2001).

\[
\langle K_{c,\text{turb,12}} \rangle = 2\pi R_{12}^2 \left\langle w_r \right\rangle \left\langle g_{12}(R_{12}) \right\rangle
\]  

Here the term of \( w_{r} \) is the radial relative velocity, which represents the “turbulent transport effect” and describes the turbulence-enhanced relative velocities of two colliding particles. The term of \( g_{12}(R_{12}) \) is the radial distribution function (RDF) at contact, which is so-called the “accumulation effect” and measures the effect of particle preferential distributions.

In this study, we propose a new collision kernel model which counts both the turbulence effects on \( w_{r} \) and \( g_{12}(R_{12}) \) (Onishi and Komori, 2006; Onishi et al., 2006). In order to develop a collision kernel model, we have carried out three-dimensional direct numerical simulations (DNSs) of particle immersed isotropic turbulent flows. Valuable data about the particle collisions were obtained from the DNSs. The data about the RDF at contact between the same size particles, \( g_{ii}(R_{ii}) \), proposed the following relations on \( R_{ii}(=g_{ii}-1) \):

\[
R_{ii} \propto \begin{cases} 
St^2 & (St \ll 1) \\
Re_{\lambda} \ St^2 & (St \approx 1), \\
Re_{\lambda}^{1/2} St^{-1/2} & (St >> 1)
\end{cases}
\]  

where \( Re_{\lambda} \) is the Reynolds number based on the Taylor microscale. These relations can be translated into the following ones;

\[
R_{ii} \propto \begin{cases} 
\left( \frac{r}{l_{\eta}} \right)^4 & (St \ll 1) \\
\left( \frac{l_{\eta}}{l_{\lambda}} \right)^2 \left( \frac{l_{\eta}}{r} \right)^2 & (St \approx 1), \\
\frac{l_{\lambda}}{r} & (St >> 1)
\end{cases}
\]  

where \( l_{\eta} \) and \( l_{\lambda} \) are the Kolmogorov scale and Taylor microscale, respectively. Equation (5) shows that \( R_{ii} \) is influenced by small scale fluid motions in the small St region, by both small and large scale motions in the region of St-1, and by large scale motions in the large St region. These are physically reasonable and therefore support the empirical relations (4). Using hyperbolic tangent functions, we can develop an empirical \( R_{ii} \) model. Fig. 1 shows examples of predictions of the developed \( R_{ii} \) model. The predictions by the model agree with the DNS results. The RDF at contact between two different size particles, \( g_{12}(R_{12}) \), can be obtained from \( g_{ii}(=R_{ii} + 1) \) using the following empirical form (Zhou et al., 2001);

\[
g_{12}(R_{12}) = 1 + \rho_{12} \left[ g_{11}(R_{11}) - 1 \right]^{1/2} \left[ g_{22}(R_{22}) - 1 \right]^{1/2},
\]  

where \( \rho_{12} \) is a coefficient, which becomes unity for \( \tau_{11} = \tau_{22} \).

Combination of our RDF model and \( w_{r} \) model proposed by Zhou et al. (2001) can lead the collision kernel model of \( <K_{c,\text{turb,12}}>. \) However, we should note that the \( <K_{c,\text{turb,12}}> \) contains only pure turbulent collisions and does not contain gravitational collisions. In order to obtain the total collision kernel, we proposed the following relation;

\[
<K_{c,\text{total,12}}> = \left( <K_{c,\text{hydr,12}}>^2 + <K_{c,\text{turb,12}}>^2 \right)^{1/2}.
\]  

Fig. 2 shows the predictions of our collision kernel model together with the DNS results in an isotropic flow with the Reynolds number based on the Taylor-microscale, \( Re_{\lambda} \), of 44. The vertical axis represents the normalized collision kernel between particles with radius \( r_1 \) (= 30 \( \mu \)m) and \( r_2 \), and the horizontal axis represents the \( r_2 \). Significant difference can be seen between the predictions of the hydrodynamic model and DNS results when \( r_2 \) is close to \( r_1 \).
30 μm), while the predictions by our model well agrees with the DNS results. This is because collisions between the same size particles are purely induced by turbulence, and the turbulence effect is not considered in the hydrodynamic model but it is included in our model.

Hereafter, we refer to the simulation based on our collision kernel model as RUN-T and the simulation based on the hydrodynamic kernel of equation (2) as RUN-NoT. The turbulence effects are considered in RUN-T, but not considered in RUN-NoT.

**Fig. 1.** Model predictions and DNS results of $R_s$ as a function of $St$.

**Fig. 2.** Collision kernels normalized by local shear rate, $λ = (ν/ε)^{1/2}$, and collision radius, $R_{12} = r_1 + r_2$, between the particles with radius 30 μm and $r_2$. 
3. Numerical Simulations of Mesoscale Convective Clouds

Fig. 3 shows the computational domain for mesoscale orographic flow over a two-dimensional mountain with the form of

$$ h(x, y) = h_0 \frac{a^2}{(x - x_m)^2 + a^2}, $$

(8)

where the standard mountain half-width $a$, the mountain height $h_0$ and the location of the mountain ridge $x_m$ were 2 km, 1 km and 20 km, respectively. The computational domain was $40 \, \text{km} \times 10 \, \text{km} \times 15 \, \text{km}$ in the $x$-(streamwise), $y$-(spanwise) and $z$-(vertical) directions and the number of the grid points were $200 \times 50 \times 75$. The height-based, terrain-following-coordinate system was chosen. Computational grids were regular in the horizontal directions and irregular in the vertical direction. The grid width in the vertical direction was set to smaller values near the bottom surface and it reached the minimum value of 10 m in the immediate vicinity of the surface.

![Fig. 3. Computational domain for orographic precipitations.](image)

The spectral bin method was implemented into the Cloud Resolving Storm Simulator, CReSS (Tsuboki and Sakakibara, 2002). A split-explicit technique was used to advance the equations in time; here we chose a time step of 0.25s for the advection terms and 0.025s time step for the fast-moving acoustic modes. We selected the horizontally explicit and vertically implicit time integration scheme for the acoustic terms and a second-order Leap-Frog scheme for the other terms. The fourth-order central difference scheme was employed for spatial derivatives. A monotonic fourth-order numerical filter was applied in the horizontal directions to suppress very short wave-length modes.

The saturated air flowed into the domain with the streamwise velocity $U = 15 \, \text{m/s}$. Initial temperature of the air flow decreased with the height at the rate of $dT/dz = 5.0 \, \text{K/km}$. The
moist Brunt-Väisälä frequency $N_m^2$ near the surface was $1.5 \times 10^5$, which indicated that the surface stratification is almost neutral. The Scorer parameter, $l = N_m/U_0$, near the surface was $2.3 \times 10^{-4}$ m$^{-1}$. Non-dimensional mountain height, $h_0l$, and half-width, $a_l$, were 0.23 and 0.46, respectively. These values were similar to those at cases A2 and A3 in Satomura et al. (2003), which reported the Steep Mountain Model Intercomparison Project (St-MIP). The results of the cases A2 and A3 showed no vertically propagating waves and the similar result has been reproduced in our simulations. The non-slip boundary condition was imposed at the bottom surface. At the upper boundary the vertical velocity was set to zero and a damping layer was used in the uppermost 5 km to avoid reflection of wave energy. In the $y$-direction, periodic boundary conditions were employed. The wave radiation condition was used to reduce the wave reflection at the outlet boundary.

In our spectral bin method, 102 classes of particle were used and the first upper boundary, $r_1$, was 5 µm and the last one, $r_{102}$, was 99 µm. The neighboring boundaries had the relation of $r_{k+1} = 1.03 \times r_k$. Particles exceeding the upper limit of radius, $r_{102}$ (= 99 µm), were redistributed into the largest class considering the mass conservations. We neglected the cold microphysics since the cloud top didn’t reach the freezing height in our simulations. The droplet collision growth was calculated based on the discretized equation of (1) using the collision kernel models listed in Table 1. The Soong’s method (Soong, 1974) was used for the condensation/evaporation term in the discretized particle equation.

<table>
<thead>
<tr>
<th>RUN</th>
<th>Collision kernel model</th>
<th>Turbulence effect</th>
</tr>
</thead>
<tbody>
<tr>
<td>RUN-NoT</td>
<td>Hydrodynamic kernel model</td>
<td>not considered</td>
</tr>
<tr>
<td>RUN-T</td>
<td>our developed model</td>
<td>considered</td>
</tr>
</tbody>
</table>

### Table 1 Summary of the collision kernel models used in numerical simulations

4. Results and Discussions

Fig. 4 shows the rain water mixing ratio at $t = 5600s$, which corresponds to a dimensionless time $t^* = Ut/\alpha$ of 42. By that time the simulations have reached a quasi-steady state. In the RUN-T, i.e., when the turbulence effect on collisions is considered, the rainfall starts from more windward position compared with RUN-NoT. This indicates that the particle collision growth is promoted by turbulence and, as a result, the precipitation timing is advanced.

Fig. 5 shows the estimated rainfall rates on the mountain surface at $t = 5600s$. In the RUN-T, the rainfall rates are larger than those in RUN-NoT at the upstream slope (-2 km $\leq x-x_m \leq 0$ km) and behind the ridge (2 km $\leq x-x_m \leq 4$ km). The spatial-averaged rainfall rate over the mountain is 1.25 mm/h in the RUN-T, which is 20% larger than 1.05 mm/h in the RUN-NoT. This means that turbulence promotes the particle collision growth and produces larger amount of rain in orographic precipitations.
Fig. 4. Rain water mixing ratio in (a) RUN-NoT and (b) RUN-T. ($t = 5600s$)

Fig. 5. Spanwise-averaged rainfall rates over the mountain. ($t = 5600s$)

5. Concluding Remarks

We investigated the effects of turbulence on cloud droplet collisions. First, we developed a turbulent collision kernel model to predict the collision frequency in cloud turbulence. Then, we developed a new spectral bin microphysics model that can be used to predict the turbulent collision growth of cloud droplets in mesoscale clouds. The effects of turbulence on cloud droplet collisions in mesoscale orographic clouds have been investigated using the developed model with sufficiently high numerical resolutions to capture the atmospheric turbulence. The simulation results have indicated that the atmospheric turbulence promotes the collision growth and increases the amount of rainfall on mountains in orographic precipitations.
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Atmospheric CO\textsubscript{2} Simulations with a High Resolution Model and Synoptic Scale Variability of CO\textsubscript{2} Column

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Abstract

We present a new version of the global atmospheric tracer transport model driven by analyzed meteorology with diurnally varying mixing in the boundary layer capable of running globally at resolutions up to quarter degree longitude-latitude or higher. The impact of the higher resolution model can be visible in resolving city plumes, airmass boundaries, diurnal cycle, fronts and synoptic scale events often observed in continuous CO\textsubscript{2} monitoring site data. To study the dynamical biases in column CO\textsubscript{2} due to possible correlation between clouds and atmospheric CO\textsubscript{2} at synoptic scale, we have made simulations of CO\textsubscript{2} (1988-2003) using NIES tracer transport model at moderate resolution (2.5 degree). Lower column CO\textsubscript{2} concentration is simulated inside cyclonic systems in summer over North hemispheric continental areas. Surface pressure is used as a proxy for dynamics and it is demonstrated that anomalies in column averaged CO\textsubscript{2} has fairly good correlation with the anomalies in surface pressure. Positive correlation, as high as 0.7, has been estimated over parts of Siberia and N. America in summer time. Our explanation is based on that the low-pressure system is associated the upward motion, which leads to lower column CO\textsubscript{2} values over these regions due to lifting of CO\textsubscript{2}-depleted summertime PBL air, and higher column CO\textsubscript{2} over source areas.

Keywords: Atmospheric CO\textsubscript{2}, Global transport model, Synoptic scale variability, Continuous observations

1. Introduction

CO\textsubscript{2} is the primary radiatively active gas and an important component of the carbon cycle. Observational data have shown that CO\textsubscript{2} levels have increased monotonically for about last half century, but at a variable rates in different years. Fossil fuel burning and deforestation are major driving force for the increase in CO\textsubscript{2}. But, several questions about its sources and sinks are remaining to be answered. Therefore, in order to have better understanding of complete process of CO\textsubscript{2} exchanges, it is important to study different processes at the higher spatial and temporal resolution using model simulations and space-based observations in addition to the ground based observations.
2. CO₂ Simulation with a High Resolution Model

Presently, there is much interest in incorporating continuous CO₂ data into inversions to estimate the CO₂ sources and sinks. Such study assumes that transport (forward) models are able to adequately simulate CO₂ concentrations at diurnal and synoptic timescales. However, this assumption may not be true as many transport models are running at coarse resolutions like 2° longitude-latitude and mesoscale variations are difficult to resolve in such resolutions.

Here we present a new version of the global atmospheric tracer transport model (NIES05) capable of running globally at the grid resolution up to 0.25° longitude-latitude. The model code is based on (Maksyutov and Inoue, 2001). The surface CO₂ fluxes used in this study are specified by Transcom continuous experiment protocol (Law et al., 2005), include seasonally varying oceanic flux (Takahashi et al., 2002), a SiB2 diurnally varying terrestrial ecosystem fluxes, and fossil fuel emissions. Higher resolution anthropogenic CO₂ emissions are synthesized from 1x1 degree emission inventory, enhanced with 2.5 min global population map data (CIESIN, 2000), and combined with lower resolution terrestrial ecosystem and oceanic flux data. The vertical resolution is enhanced to 47 levels for better resolving the mixing processes in the boundary layer, driven by 3-hourly PBL height data of ECMWF. Four types of horizontal resolution (2x2, 1x1, 0.5x0.5 and 0.25x0.25) have been tested.

Fig. 1 shows the surface CO₂ distributions in East Asia at the resolutions of (a) 2x2 and (b) 0.25x0.25 degree. Although the area-average concentration values are almost the same in Fig. 1(a) and (b), distribution shapes are quite different from each other. For example, 0.25x0.25 model result (Fig. 1b) shows more vivid city plumes than from 2x2 model result (Fig. 1a). It also shows clearer vortex shape due to a typhoon near Kyusyu.

![Fig. 1. Surface CO₂ distributions over East Asia on 30August, 2002 at 0300 hour using (a) 2° x 2° resolution and (b) 0.25° x 0.25° resolution.](image)

Higher resolution model shows better results at Tsukuba (Fig. 2a), particularly during the 4th to 7th August period when winds from Tokyo dominate. Fig. 2b shows results at Pt. Barrow. In general, synoptic scale variabilities have been captured by model at all resolution, but higher resolution results show better comparison with observations.
Fig. 2. Surface CO₂ distributions over (a) Tsukuba and (b) Pt. Barrow. Results of model calculations with various resolutions and observation.

3. Column CO₂ and Low Pressure System

CO₂ column measurements are potentially important input for atmospheric CO₂ inversion calculations because of much lower contributions from errors in modeled vertical convection. These column measurements from space-borne sensors could be advantageous over ground-based measurements, however one of the major challenges is the requirement of measurements with very high precision (better than 1%) in space-borne CO₂ observations. Simulation shows that precisions of ~0.3–2.5 ppmv can be achieved, but using only for clear sky conditions data (Kuang et al., 2002). Observations in cloudy condition are fraught with difficulties and generate greater uncertainty in most of the space-borne observations. On average cloud cover is 50–60% on annual basis and sampling only in clear-sky conditions might have biases, as photosynthesis is stronger in clear-sky condition in many ecosystems and thus measured CO₂ could be lower than the average values.

Analysis is performed to study the influences of synoptic scale weather systems on the column averaged CO₂. Surface pressure is used as a proxy for dynamics and dynamical biases in column CO₂ using the correlation between column averaged CO₂ anomalies and surface pressure anomalies are estimated. Long-term (1988-2003) CO₂ simulation is made using NIES tracer transport model (2.5°x2.5°) and fluxes for fossil fuels, terrestrial biosphere (CASA NEP), the ocean, and inverse model derived (using SVD truncation) regional fluxes (11 land and 11 ocean regions) are used. Fig. 3 shows variations in normalized column averaged CO₂ mixing ratios and surface pressure over a part of Siberian region. Lower CO₂ values associated with a low-pressure system can be seen here. It is estimated that ninety five percent probability range of CO₂ anomalies is from -0.3 to 0.81 ppmv for the Northern Hemisphere, while this range is larger (-0.45 to 1.2 ppmv) for the part of Siberian region.
Fig. 3. Variations in normalized column average CO₂ mixing ratios in ppmv (color shaded contours) and surface pressure in hPa (line contours) over a part of Siberian region. Low CO₂ values are associated with a low-pressure system.

Linear Pearson correlation coefficient between anomalies in column CO₂ and surface pressure is estimated for each grid point on monthly bases. A covariance matrix is also generated. Generally average correlation coefficient varies from -0.6 to 0.6 over different regions on yearly basis. During July and August correlation coefficient is as high as 0.8 over part of the Siberia. Significant positive correlation is also observed over part of N. America. It appears that surface pressure explains a significant part of the variance in CO₂. Fig. 4 shows positive and negative biases in derived CO₂ due to varying pressure systems in August for the period 1988-2003. It is shown that CO₂ mixing ratio could have a positive bias of more than 0.5 ppmv with the varying pressures over Siberia. Summertime low-pressure system is associated the upward motion, which leads to lower column CO₂ values over these regions due to lifting of CO₂-depleted summertime PBL air. Therefore in the cloudy condition, when low-pressure system prevails, CO₂ mixing ratios could be lower by this value. Fig. 2(b) shows monthly variations in the CO₂ bias over a part of Siberia during different years. Biases are higher than 0.8 ppmv during July, August, and September in some years. Generally uncertainty in these estimated is less than 0.4 ppmv. Therefore it appears to be essential to consider these biases while analyzing the column CO₂ data from space-borne observations, because most optical observations are not available under cloudy conditions typical for the low-pressure systems.
Fig. 4. (a) Biases in CO₂ mixing ratios due to low-pressure system during August for the period 1988-2003. Positive values for example, suggest that CO₂ mixing ratio could be low in case of a low-pressure system, while negative values suggest for higher mixing ratios in low-pressure system. (b) Monthly variations in CO₂ biases over a part of Siberian region (75° - 120°E, 58° - 75°N) in different years of 1988-2003 period.

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Numerical Experiment on the Interaction between Large-Scale Atmospheric Motion and Cumulus Convection: Mechanism of Spontaneous Large-Scale Stationary Concentration of Cloud Activity

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Abstract
We performed sensitivity experiments to examine the appearance of the very large-scale stationary concentration of cloud activity that occurs spontaneously in the two-dimensional cloud-resolving model. The model extends 32,768 km in the horizontal direction and is integrated 60 days with surface fluxes from the ocean and a constant body cooling in the model troposphere being applied. Two experiments are reported with different specifications of the vertical profile of body cooling. The results suggest that the "planetary-scale" stationary modulation appear as long as propagating disturbance of several thousand km scale develops. The coupling between cloud activity and the large-scale motion is examined for various wavelength and is discussed with the predictions of linear wave-CISK theory in mind. We also report briefly on our subsequent efforts to develop the new cloud-resolving model.

Keywords: Cumulus convection, Wave-CISK

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.

Nakajima (1994) developed a two-dimensional cloud model that covers a domain of 16,384 km where the interactions between cumulus convection and the large-scale motions to be represented explicitly, and found that several types of large-scale cloud organizations develops spontaneously in the long-term integrations of the model, some of which are related to wavy conditional instability of the second kind (wave-CISK) and the propagating instability with wind-induced surface heat exchange (WISHE). As an extension of Nakajima (1994), Nakajima (2002) 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. Nakajima (2003) examined the former type of organization of clouds and interpreted it as a new type of wave-CISK. Nakajima (2004), based on numerical
experiments extended up to 160 days, focused on the intra-seasonal time scale vacillation of domain averaged convective activity. Nakajima (2005) examined the space-time structure of cloud convection in a 32,768km domain with the setup where WISHE feedback is deactivated and found that, in addition to the propagating pattern governed by wave-CISK, a very large scale stationary modulation of cloud activity slowly develops and amplifies to result in the appearance of one concentrated region of intense cloud activity. Nakajima et al. (2006) conducted a series of sensitivity experiments and found that the amplitude and the horizontal scale of such stationary region of cloud convection depend on the intensity of the body cooling in the troposphere and the strength of the coupling between the atmosphere and the ocean. Namely, more than one regions of concentrated cloud activity can develop in the computational domain when both the body cooling and the atmosphere-ocean coupling are strong.

The present paper examines the physical mechanism that results in the development of the stationary large-scale cloud concentration. Nakajima (2005) suggested two possible candidates: large-scale convective instability (CIFK) and a modification of wave-CISK. In this paper, the latter possibility is pursued through the analysis of the numerical experiment. Also, we repeat the numerical experiment of Nakajima (2002), which showed the appearance of wave-CISK in a realistic body cooling profile, in an extended computational domain, where large-scale stationary concentration of cloud activity is found to develop, as will be shown later. Finally, we briefly mention our subsequent efforts for the development of an updated numerical model that is more flexible and suitable to be executed on parallel computer.

2. Mechanism of Stationary Concentration of Cloud Activity

2.1 Design of Experiments

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.

The vertical profile of the troposphere cooling is specified to be top heavy, with which the stationary large-scale concentration of cloud activity developed in Nakajima (2005). The wind speed in the surface flux calculations is fixed at a constant, homogeneous value in each experiment. By that, any WISHE feedback will be switched-off. This experiment will be called “case H”.

We perform another experiment where the body cooling is strong in the lower and the upper levels and is weak in the middle levels. This experiment is basically same as that reported in Nakajima (2002) except that the computational domain is larger and the wind speed for surface flux calculation \( V_{sk} \) is increased to 30m/s. This experiment will be called “case HL”.

2.2 Sensitivity of the Structure Cloud Activity to the Vertical Structure of Cooling

Fig. 1 shows the temporal evolution of rainfall intensity in case H. In the earlier period,
the structure of precipitation is dominated by the right- and left-propagating pattern. This is a manifestation of wave-CISK. However, a large-scale modulation with wave number one gradually develops, and all of the cloud activity occurs at one concentrated area after 40 days of time integration. It is clear that a large-scale modulation of cloud activity can amplify in this setup. Fig. 2 shows the evolution of rainfall intensity in case HL. As was reported for a similar experiment in Nakajima (2002), propagating organization of convective activity develops in the early periods of the experiment. In later time, however, a stationary modulation with wavenumber one emerges and develops. The gross feature of cloud activity, i.e., the co-existence of the propagating disturbance of several thousand kilometers scale and the stationary wavenumber one modulation, is common to that found in case H. This character also appears in experiments with various values of $V_{sc}$.

The similarity of the spatial and temporal structure of cloud activities in case H and case HL suggests that the propagating and the stationary organizations of clouds have a common mechanism. More specifically, as was suggested by Nakajima (2005), wave-CISK is responsible not only for the propagating disturbance but also for the stationary large-scale modulation.

Fig. 1. Temporal evolution of rainfall distribution in case H.

Fig. 2. Temporal evolution of rainfall distribution in case HL.
2.3 Dependence of Heating Parameter on the Scale of Disturbance

As was suggested in Nakajima (2005) based on linear stability analysis of wave-CISK, both propagating and stationary unstable mode can develop for the same vertical structure of cumulus heating, depending on the strength of the coupling between the cumulus heating and large scale vertical motion; the stationary unstable mode appears for strong coupling, and the propagating mode appears for moderate coupling. The results of the experiments presented here imply that the strength of the coupling may vary depending on the horizontal scale of the disturbance. Following this line, we examine the correlations of the horizontal Fourier components of the low level vertical motion and those of precipitation for various wavelengths.

Fig. 3 shows the dependence of the proportional coefficient of the Fourier component of the precipitation to that of the low level vertical motion on wavenumber in case H. The same quantities for the case with bottom heavy cooling, where neither the propagating disturbance nor the stationary large-scale modulation develops, are shown. It is evident that the coupling between the precipitation and vertical motion strongly depends on the horizontal scale; the intensity of the coupling at wavenumber one, corresponding to the stationary modulation, is almost five times as large as that for wavenumbers around 10, which correspond to the horizontal scale of the propagating structure. Bearing the prediction of wave-CISK theory mentioned above in mind, this significant change of the strength of interaction between cumulus and large-scale motion over horizontal wavelength space is consistent with the space-time structure of cumulus activity in case H and case HL. That is, the disturbance with wavenumber one represents stationary growth because the coupling is strong, and, on the other hand, the disturbance with wavenumbers around ten represents propagating growth because the coupling is moderate.

Fig. 3. Dependence of the wave-CISK parameter on horizontal wavenumber in case H (green squares) and the case with bottom heavy cooling (red dots).
2.4 Possible Selection Rule for the Emergence of Large-Scale Stationary Model

Following this line of suggestion, one may suppose that, in case with bottom heavy cooling, stationary growth of large-scale modulation did not appear because the coupling is weaker for larger-scale disturbance. However, this is no true; as is also shown in Fig. 3, the coupling is stronger for larger scale also in case with bottom heavy cooling. Therefore, the enhancement of the coupling between clouds and vertical velocity in larger scale is not sufficient factor to the development of the large-scale stationary modes.

A possible “missing link” may be the behavior of unstable modes in “intermediate” scale. For the case H and HL, assuming that the cumulus heating balances with the cooling, wave-CISK analysis predicts the existence of propagating unstable modes for “intermediate” wavelength where the coupling is moderate. On the other hand, for the case with bottom heavy cooling, wave-CISK analysis predicts stationary unstable modes for all wavelengths. It may be true that the propagating unstable modes in case H and HL “smooth out” the cloud activity in the domain, effectively preparing a uniform environment for the wavenumber one stationary mode. This kind of “smoother” does not exist in the case with low level enhanced cooling, and this may result in the absence of significant larger-scale stationary modes. Of course, this is only a speculation, and requires harder evidence, which may be obtained by the future study.

3. Development of an Updated Model: Sequel

As was written in Nakajima et al. (2006), we are developing an updated model of moist employing the quasi-compressible system. The model is now successfully coded and is tested on the SX-6 of CGER. The preliminary results suggest the trade-off between the readability (or, actually, the “writability”) of the codes in fortran 90 and the performance on the SX-6 is rather serious. For a typical example, the new model runs about five times slowly than the old model written in conventional fortran 77 style. Based on the diagnostic output of the test run, main difficulty seems to exist not only in the vectorization but also in the optimization; the number of instruction per step is about two times larger for the new code. In spite of the handicaps in efficiency, the new model will be much superior in parallel computing environment. We are planning to explore this point in future study.

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3. Geophysical Fluid Dynamics
Differential Diffusion of Heat and Salt

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Abstract
Vertical diffusivity coefficients of salt and heat in the doubly stratified water/ocean are computed. The results show that they differ significantly due to the difference in the molecular diffusivity. The difference is conspicuous in the time-averaged value since the initial difference dominates. These results show that the diffusivity coefficients of salt and heat might have large difference as have been suggested by recent studies. If the initial time development is important, dependence on the initial conditions would be large, meaning that the theoretical prediction might be possible.

Keywords: Differential diffusion, Salt, Heat, Ocean, Turbulence

1. Introduction

Ocean water has a density variation in the vertical direction, i.e., it is density stratified vertically, due to both the temperature and the salt distribution. Therefore, its motion is affected by the unstable fluid motion due to “double-diffusive” instability, which arises from the difference in the molecular diffusivity between heat/temperature and salt (Stommel et al., 1956). Therefore, most of the previous studies treat intrinsically unstable state, where either the temperature or the salinity is unstably stratified, meaning that water with lower temperature or larger salinity exists above the fluid layer with higher temperature or smaller salinity.

However, recent studies in oceanography show that “differential” diffusion, i.e., difference in diffusion of heat and salt, occurs even in a completely stable fluid, which is stable both in temperature and salinity. Therefore, the differential diffusion has become one of the vital problems in the ocean global/local circulation model, and The International Association for the Physical Sciences of the Oceans (IAPSO) has organized a special working group on this problem (Gargett 2003, Gargett et al., 2003). At present, ocean circulation models usually use the same eddy diffusivity coefficient for the heat and the salinity, on the assumption that the ocean circulation velocity (Reynolds number) is sufficiently large and nonlinear effects dominate the fluid motion to overcome the molecular diffusion. However, the Reynolds number of the oceanic flow is often small actually, and the assumption is not satisfied. Indeed, recent ocean observations show growing evidence that there is a difference between the heat flux and the salt flux (Nash and Moum 2002), which means the difference between the eddy diffusivity coefficients.

In addition, conventional theories predict only the initial transient stability or instability of specific length scale of the fluid motion, and they do not predict the “net” integrated effects of those stabilities/instabilities which appear as heat fluxes or salt fluxes. The fluxes eventually determine the distribution of temperature and salt concentration in the ocean, and they must be determined either by numerical simulations or by another new theory, which can predict the time-integrated effects of the fluxes.
In this study we consider a system with two density-stratifying substances such as salt and heat, and obtain the numerical evidence that the eddy diffusivity coefficients of salt and heat should be different. We assume that the two substances are both stably stratifying, so that the differential diffusion occurs while the double-diffusive instability does not occur.

We consider decaying turbulence which is calculated by the spectral method at the resolution of $128^3$ or $256^3$. Since the computation of scalars with small molecular diffusivity, such as heat ($Pr = 6$) or salt ($Pr = 600$), requires vast computational resources since they have very small scale structures compared to the velocity, we use here $Pr = 1$ for the heat and $Pr = 6$ for the salt, to illustrate preliminary the qualitative effects of differential diffusion. The results indicate that the eddy diffusivity coefficient is usually much smaller for salt (with smaller molecular diffusivity) than heat (with larger molecular diffusivity).

2. Flow Structures and the Difference between the Heat Flux and the Salt Flux

In Fig. 1 we show the iso-surface of the vorticity magnitude, characterizing the flow in differential diffusion. In fact, the flow structure itself is not very different from the case of single stratification, i.e., stratification only by heat or salt. The flow contains pan-cake structures, which appears also in the single stratification. However, the structure depends on the Reynolds number $Re$, and with higher Reynolds number /or with higher fluid velocity, the structure is composed of thinner pancakes.

![Iso-surface of vorticity](image)

Fig. 1. Iso-surface of vorticity $|\omega|$ at $Re_\lambda = 70$, $Fr = 0.4$, $Pr_T = 1$ and $Pr_S = 6$. 
Fig. 2. Iso-surface of vorticity $|\omega|$ at higher Reynolds number with $Re_\lambda = 100$, $Fr = 0.4$, $Pr_t = 1$ and $Pr_S = 6$. The vortex shape is thinner at this higher Reynolds number.

Fig. 2 shows the flow structure at a higher Reynolds number of $Re_\lambda = 100$, showing the existence of thinner pancakes. On the other hand, at lower Reynolds numbers ($Re_\lambda < 50$) pancake structures do not appear, showing different flow structures at small scales.

In Fig. 3 we show the time development of the kinetic energy ($KE$) due to velocity fluctuation, potential energy ($PE$) due to temperature and salinity fluctuation, and the total turbulence energy ($KE + PE$). Here, time $t$ is normalized by the buoyancy time, i.e., $t_N = Nt/2 \pi$. If we look at the red lines which represent the case in which the stratification/buoyancy force is applied from the initial instant, there is a time oscillation in the kinetic energy $KE$ and potential energy $PE$ due to the energy exchange between them, while the total turbulence energy $KE + PE$ decays monotonically. The results shown by blue lines are those of different initialization method, but they give similar behavior. In these results, initial potential energy is set to be zero, so that it increases initially until it reaches an equilibrium value which balances with the kinetic energy.

Fig. 4 shows the time development of the vertical temperature flux ($Pr = 1$) and vertical salt flux ($Pr = 6$) given respectively by

$$K_T = -\frac{T_w}{dz} \quad \text{and} \quad K_S = -\frac{S_w}{dz},$$

where $w$ is the vertical velocity. These are the same as the eddy diffusivity coefficients usually used in the ocean circulation models. We note that, except for the very initial time development, $K_S$ is smaller than $K_T$, although the difference decreases as both fluxes decreases
with time. This suggests that the initial difference persists for an indefinitely long time if we consider the time-integrated effect, which would be necessary for the ocean modeling where long-time scale phenomena are usually important.

![Graph](image)

**Fig. 3.** Time development of the kinetic energy $KE$ due to velocity fluctuation, potential energy $PE$ due to temperature and salinity fluctuation and the total turbulence energy $KE+PE$. Red lines correspond to the case when the stratification is applied initially (at $t = 0$), while blue lines correspond to the case when the stratification is applied at $t = 1/u$, i.e. after the nonlinear interaction of turbulent eddies has been well developed.

![Graph](image)

**Fig. 4.** Time development of the eddy diffusivity coefficient of temperature ($K_T$) and salt ($K_S$).
Fig. 5. Time development of the time-integrated eddy diffusivity coefficients of temperature ($\Phi_T = \int_0^{t_N} K_T dt$) and salt ($\Phi_S = \int_0^{t_N} K_S dt$).

Fig. 6. Time development of the ratio $d$ of the time-averaged eddy diffusivities, i.e., $d = \Phi_S / \Phi_T$.

Fig. 5 shows the time development of time-integrated eddy diffusivity coefficients of temperature and salt are defined by

$$\Phi_T = \int_0^{t_N} K_T dt \quad \text{and} \quad \Phi_S = \int_0^{t_N} K_S dt.$$  

We note that they both approach constant values as $K_T$ and $K_S$ both decreases. They reach asymptotic values at about $t_N = Nt/2\pi \sim 2$, showing the importance of initial time development. On the other hand, this suggests that if we could predict the initial time development well, we can predict the time-averaged value of the eddy diffusion coefficient.

Fig. 6 shows the time development of the ratio between the time-averaged eddy diffusivity coefficients, i.e.,
\[ d = \frac{\Phi_S}{\Phi_T} = \frac{\int_0^t K_S dt}{\int_0^t K_T dt}. \]

If this value is significantly different from 1, we note that the present values of the eddy diffusivity coefficients used in the ocean circulation model \((d = K_S / K_T = 1)\) might contain large errors. In the present case, the value is about 0.5–0.6, suggesting a significant difference between \(K_T\) and \(K_S\) which should be used in the ocean circulation models.

3. Conclusions

We have simulated the differential diffusion of heat \((Pr = 1)\) and salt \((Pr = 6)\) in the fluid doubly stratified by the heat and salt. Time development of the vertical heat and salt fluxes, or the eddy diffusivity coefficients, showed significant difference between heat and salt, and their ratio reduces to as low as 0.5, although the difference would depend on the buoyancy Reynolds number as demonstrated by laboratory experiments (Jackson and Rehmann 2003), and further investigation is necessary to know what determines the value of \(K_S / K_T\) which should be used in the ocean circulation models.

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Equatorial Precipitation Patterns in Aqua-Planet Experiments: Effects of Vertical Turbulent Mixing Processes

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Abstract
An aqua-planet general circulation model (GCM) is utilized to examine the effects of implementation of the vertical turbulent mixing and surface flux processes. When the value of bulk coefficient of the surface flux formulation is fixed and the vertical turbulent mixing process is replaced by a simple diffusion with a fixed coefficient of a given profile, there appears multiple peaks of grid scale precipitation propagating eastward along the equator. This indicates that the feedback of the variation of circulation field to the turbulent processes has a significant effect in the appearance of precipitation patterns. The difference of the implementation of turbulent processes may explain the difference of the appearance of equatorial precipitation patterns from those of the earlier studies with the same cumulus parameterization scheme. While performing the aqua-planet experiment with the old AGCM5, we have implemented such physical processes as cumulus convection and turbulent mixing into our new GCM, which is supposed to enable us to perform experiments like above more easily, and performed some verification experiments by comparing the results with those of the older one. In addition, we have provided an additional function for RDQC in order to produce the reference manual of programs automatically from the source codes.

Keywords: GCM, Aqua planet experiment, Super cloud cluster, Equatorial disturbances, Vertical turbulent mixing

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 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. The other 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 basic numerical experiments mentioned as our first subject are also expected to contribute to providing guidelines for this model construction.

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 methods of cumulus parameterization. However, we still do not understand whether precipitation activities represented by the model without resolving the fine structures of cumulus activity are really reasonable. Actually, in recent years, an international aqua-planet comparison project, APE (http://www-pcmdi.lanl.gov/projects/amip/ape/) has been carried out to investigate the variety of the representations of precipitation and circulation characteristics obtained in the major climate and/or numerical weather prediction models under the condition of aqua-planet. However, at the moment the analyses of the APE results have been carried out only a little. One of the reasons for this is that we have some knowledge about the diversity of calculation results as reported last year, however, we have little knowledge about the cause of the diversity. We cannot at the moment figure out where to confine our attention to classify the diversity of calculation results. With those situations in mind, we are now trying to carry out a series of switch-on experiments of physical processes such as cumulus convection and turbulent processes. What we found here is that, not only the cumulus convection process, the implementation of turbulent mixing processes has great influence on the appearance of precipitation patterns. In section 2, we describe our recent results of the switch-on experiments of physical processes.

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. This year, we started to implement some physical processes into our new primitive equation model on a sphere. In due course, we have provided an additional function for RDoc in order to produce the reference manual of programs automatically from the source codes. In section 3, we describe our present status on the development of our new primitive equation model, Dennou-Club Planetary Atmospheric Model (DCPAM), and the development tool related to DCPAM development.

2. Dependence of Equatorial Disturbances on the Vertical Cooling Profile

2.1 Purpose

For the purpose of approaching the problem of diversity of precipitation patterns, we carried out experiments with various vertical profiles of radiative cooling by the use of aqua
planet GCM with simplified physical processes (e.g., Hayashi et al., 2005). We demonstrated that, in the experiment with Kuo scheme as the cumulus parameterization, the run with the radiative cooling maximum located in the upper troposphere (referred to as run K-UC, hereafter) yields eastward migration of the grid scale precipitation areas. The composite structure of the eastward precipitation features of K-UC run shows an upward-westward tilt of phases which is consistent with the structure expected by the wave-CISK theory as in Numaguti and Hayashi (1991a).

However, the characteristics of the appearance of eastward migrating precipitation features in our experiments is somewhat different from those in the earlier studies of Hayashi and Sumi (1986) and Numaguti and Hayashi (1991a) where the same Kuo scheme is utilized as the cumulus parameterization scheme. The major difference is that almost only one grid scale precipitation area appears in our present experiments, while many coherent grid scale precipitation areas appear along the equator in the earlier studies. Actually, one of the guesses about the results of the radiation profile experiment was that the wave-CISK like process might operates effectively when the heating peak is located at the upper troposphere so that a number of coherently eastward moving precipitation areas might appear along the equator. We suspected the reason for the different appearance of our recent results with the older ones was that the model representation of radiation and/or cumulus parameterization is different. What actually happened is that the wave-CISK like process does operate but produces only one intense convection area along the equator. We could not produce multiple precipitation peaks along the equatorial circumference by changing the vertical profile of radiative cooling.

Now, another possible cause of the difference of the appearance of precipitation patterns is the difference of the implementation of the turbulent processes, such as the surface flux formulation and the turbulent vertical mixing. They greatly contribute to the moisture distribution especially when the cumulus process does not operate, and hence may have a great influence on the appearance of precipitation patterns. Naturally, there is a large gap between the setup of our GCM and that of the simple linear wave-CISK model. In order to relate the calculation results of our GCM with the simple wave-CISK theory, it is desired to prepare a hierarchy of models with the various complexities of processes from the linear model to the full non-linear model with water budget and subgrid turbulent parameterizations. In the followings, we will present several results obtained by simplifying the physical processes of our GCM, especially the vertical turbulent processes.

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. 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. The vertical mixing process is represented by Mellor and Yamada (1974) level 2, and the surface flux is evaluated by the bulk method of Louis (1979) and Louis et al. (1982).

We have performed the following runs on top of the K-UC run of Hayashi et al. (2005).
• Control run, K-UC of Hayashi et al. (2005): the radiative cooling profile has its peak at the upper troposphere so that coherent eastward motion of precipitation areas along the equator is clearly observed as is demonstrated in Fig. 1.
• Fixed radiative cooling profile run: the vertical profile of radiative cooling is given as the time and zonal mean of the control run.
• Fixed turbulent coefficient run: in addition to the setup of fixed radiative cooling profile experiment, the vertical profile of the vertical mixing coefficient is given as the time and zonal mean of the control run, and the bulk coefficient and the magnitude of surface wind for Louis formula are also given as the time and zonal mean.
• No turbulent process run: in addition to the setup of fixed radiative cooling profile experiment, the mixing process by Mellor and Yamada and the surface flux by Louis are switched off.

The period of integration is 40 days starting from a certain time of K-UC experiment as the initial condition.

2.3 Results

The important characteristic of the results of the control run, K-UC, is that there appears almost only one intense precipitation feature at the equator which propagates eastward coherently (Fig. 1). The associated temperature anomaly has a rather wide longitudinal extent and is dominated by wavenumber one signal. To the west of the intense precipitation, there appears cold anomaly in the lower troposphere, while to the east, there appears intense warm anomaly in the lower troposphere.

The results of the fixed radiative cooling profile run (not shown here) show little difference compared to the results of the control run. It seems that the circulation anomaly fields associated with the intense precipitation are even intensified and become clearer. However, this could be within the range of the fluctuation of the disturbance.

As for the fixed turbulent coefficient run, the precipitation and circulation features along the equator are significantly modified from those of the control run. Since the initial condition is the same as the control run, the eastward motion of wavenumber one structure can be seen at the early stage of the time development, especially, of the temperature field (Fig. 2a lower panel). However, at around 10 days, there appear a few narrow areas of warm temperature anomaly which propagate coherently eastward. Correspondingly, in the snapshot at day 39.8, there appear two or three additional warm temperature anomalous regions and also cool temperature anomalous regions in the troposphere (Fig. 2b lower panel). It might be difficult to observe in the snapshot figure of precipitation (Fig. 2b upper panel), but the number of intense precipitation areas is increased, and their coherent eastward motion can be observed in the time development of precipitation along the equator (Fig. 2a upper panel).

It seems that there is a significant difference in the field of humidity between the fixed turbulent coefficient run and the control run (Fig. 1 and 2 middle panels). The intensity of dryness and wetness of the dry and wet regions of the fixed turbulent coefficient run is weak. The weak contrast of moisture corresponds to the increase of the number of precipitation areas. Cumulus convection occurs more easily when the dryness of the atmosphere above the surface boundary layer is weak and also the dryness in the mixed layer is weak. Since the number of precipitation increases, the accumulation of humidity is less intense in the mixed layer; the anomalous accumulation of humidity in the mixed layer is smoothly released by convection.

As for the no turbulent process run (Fig. 3), the eastward propagating features are gradually destroyed, and at about day 20, westward propagating features emerge. The
expectation before the experiment was that the wave-CISK like mechanism did not suffer from the disappearance of the turbulent processes, and hence the eastward moving features were sustained. The calculation results seem to deny this expectation. The eastward motion of circulation fields remain to a certain extent, but the behavior of moisture is obviously decoupled from the eastward motion of the circulation field.

Fig. 1. Control run, K-UC. (a) Temporal variation at the equator for the period of 40 days: (upper) precipitation (kg m$^{-2}$ s$^{-1}$), (middle) specific humidity (kg kg$^{-1}$) at the lower troposphere $\sigma=0.83$, (lower) temperature (K) at the middle troposphere $\sigma=0.55$. (b) Longitude and height cross sections along the equator at day 39.8: (upper) condensation heating (K s$^{-1}$), (middle) specific humidity (kg kg$^{-1}$), (lower) temperature (K).

Fig. 2. Same as Fig. 1 but for the fixed turbulent coefficient run.

Fig. 3. Same as Fig. 1 but for the no turbulent process run.
2.4 Summary

Numaguti and Hayashi (1991b) also calculated the case with the fixed bulk coefficient and surface velocity in the surface flux evaluation of the Louis formula. The tendency that the wavenumber one structure is weakened is also observed in their results. They concluded that the wavenumber one feature is produced by wind induced surface heat exchange (WISHE). However, it seems that, whatever the reason of the moisture contrast is, if there is a certain feedback process which enhances the moisture contrast, then the wavenumber one structure appears. It seems that the feedback caused by wind shear and/or vertical stability implemented in the Louis formula have also a certain importance in producing the moisture contrast. Moreover, it is worth considering the moisture contrast in the free atmosphere above the mixed layer. The enhancement of the moisture contrast in the troposphere has obviously a significant effect in the appearance of precipitation features. We have performed rather ad-hoc preliminary calculations to observe the possible sensitivity of the moisture contrast caused by the turbulent mixing processes. A more systematic investigation of the effects of turbulent processes is necessary to clarify what is going on in the model and relate those results with what we know from the simple wave-CISK model. These are to be carried out now and to be reported in the near future.

3. Toward a Hierarchical Set of Models from Simple to GCM: DCPAM

We have been developing, as a project of GFD Dennou Club, a hierarchical set of models covering from simple standard experiments for geophysical fluid dynamics, such as wave-CISK, to experiments for climate of planetary atmospheres. DCPAM is a general circulation model for experiments of earth and planetary atmospheres for which readability and modifiability of the source codes are carefully considered in order to enable us to perform such experiments as changing physical processes like described in section 2 more easily. Last year, we developed the dynamical core of the model and performed Held and Suarez (1994) test. This year, we have implemented a few physical processes to perform the same aqua planet experiment described so far.

The physical processes implemented this year are from our AGCM5.3 which is utilized in the previous section. They are dry and moist convective adjustments, large scale condensation, radiation consisting of four bands with artificial absorption coefficients, vertical turbulent mixing by Yamada and Mellor level 2, and the surface flux of the bulk method by Louis. The dynamical core is the same as that developed last year, that is, the spectral transform method is utilized for the horizontal discretization and the Arakawa and Suarez (1983) scheme is adopted for the vertical discretization.

A 100 day integration of aqua planet configuration with fixed SST distribution is performed with explicit time integration scheme and the time filter of Asselin (1972). The efficiency of calculation per one step with the resolution of T21L16 is about 1/2.3 compared to that of our AGCM5.3. We are now re-examining the programming style of the source codes in order to improve the execution performance of the model, while keeping the readability and modifiability of the codes.

While constructing the model source codes, preparation of reference manuals which describe the modules, subroutines and functions of the model is increasingly necessitated. Those reference manuals are essential not only for the end users but also for the developers to enable smooth exchange of the detailed information on the separately built program units. In our source code, the description statements of the program units are embedded in the program
source codes as Fortran comments with some simple control codes, so that the reference manuals are automatically produced from the source codes through the interpreter, RDoc. RDoc is a text interpreter written by Ruby, an object oriented script language, to produce HTML documents. RDoc is originally designed for producing HTML reference manuals of Ruby programs by interpreting the comment lines of the Ruby source codes. RDoc were improved to interpret C and Fortran90, while the functions for Fortran90 were insufficient for large scale programs like ours. We have further improved the functions of RDoc to interpret Fortran90 programs to produce reference manuals automatically in the form of HTML; the arguments are analyzed with the help of the comment lines, links to the reference manuals of the program units utilized, and links to the source codes are automatically produced. The resultant manual page for DCPAM is now seen from http://www.gfd-dennou.org/arch/dcpam/dcpam3/dcpam3_current/doc/code_reference/htm/.

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On Similarity between Momentum Transfer and Scalar Transfer in the Airflow over Traveling Waves

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Abstract
We investigated the similarity between the momentum transfer and the scalar transfer in the airflow over traveling waves by carrying out the direct numerical simulations. The simulations for 10 wave age were carried out, and we investigated the wave-age dependence of the momentum transfer and the scalar transfer. The results showed that wave-age dependence of the non-dimensional roughness height for the momentum transfer was similar to that for the scalar transfer. This indicates that there is a similarity between the momentum transfer and the scalar transfer in the airflow over a traveling wave train.

Keywords: Direct numerical simulation, Air-sea interaction, Turbulent transfer

1. Introduction

Sensible heat and latent heat transferred from the sea to the atmosphere are the energy source of tropical cyclone. Therefore, it is important to comprehend their exchange processes and estimate their exchange rate precisely, when we predict the atmospheric and ocean phenomena numerically.

The exchange rates of the sensible heat and latent heat have been measured by some field observations using the towers or the ships. In most of the parameterizations of the heat fluxes proposed by carrying out those observations, the heat exchange coefficient is constant with wind speed (e.g., Smith, 1988; Anderson, 1993; DeCosmo et al., 1996). However, it is difficult to understand the characteristics of the exchange processes in detail by analyzing these datasets, since these datasets are significantly scattered due to the effects of the atmospheric stability and the sea surface state.

Using the datasets measured at the coastal zone, Mahrt et al. (1998) discussed the dependence of the sensible heat exchange coefficient on the wave age, which is the characteristic parameter representing the sea surface state. They showed that the heat exchange coefficient decreases as the wave age increases, and the dependence is the same as that of the momentum transfer coefficient (drag coefficient). Their result indicates that the wind wave affects the heat transfer in the airflow. However, there are few studies investigating the effect of the wind waves on the heat exchange processes, so that the knowledge remains to be poor.

In the present study, carrying out the direct numerical simulations (DNS) of the airflow over a traveling wave train, we investigate the wave-age dependence of both the momentum
transfer and the passive scalar transfer, and discuss the relations between them. We calculate the transfer of passive scalar and we do not consider the effect of buoyancy. The simulations for 10 wave ages \( c/u \), \( (c/u, = 0, 2, 4, 5, 6, 8, 10, 12, 16 \) and \( 20) \) are carried out, where \( c \) is the wave velocity and \( u \) is the friction velocity of air.

2. Direct Numerical Simulation

The flow considered in the present study is bounded by a bottom two-dimensional traveling wave train and an upper free-slip wall, as shown in Fig. 1. Periodic boundary conditions are assumed in the horizontal directions. In the frame of reference moving with the traveling waves, the shape of traveling wave train is invariant with time. The flow is driven by a constant horizontal pressure gradient, and we carry out the simulation till the flow reaches the fully developed and statistically steady state. The governing equations are continuity equation and Navier-Stokes equation for incompressible fluid, and the passive scalar conservation equation:

\[
\frac{\partial u_i}{\partial x_i} = 0, \tag{1}
\]

\[
\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} = -\frac{1}{\rho_a} \frac{\partial p}{\partial x_i} + \nu \frac{\partial^2 u_i}{\partial x_j^2} - \frac{\hat{u}^2}{h} \delta_{ij}, \tag{2}
\]

\[
\frac{\partial T}{\partial t} + u_j \frac{\partial T}{\partial x_j} = D \frac{\partial^2 T}{\partial x_j^2}, \tag{3}
\]

where subscript \( i (=1,2,3) \) denotes the streamwise \( x \) (\( x_i \)), spanwise \( y \) (\( y_i \)) and vertical direction \( z \) (\( z_i \)). The origin of \( z \) (\( z = 0 \)) is taken on the mean water level. The variables \( p \), \( \nu \), \( T \), \( D \) and \( (u_1,u_2,u_3) = (u,v,w) \) denote the pressure, the kinematic viscosity of air, the passive scalar, the molecular diffusivity of the passive scalar, and the velocity in the \( (x,y,z) \) direction. The last term of the right-hand side of (2) means the constant streamwise pressure gradient. In the present study, we carried out the simulation with a fixed value of virtual friction velocity \( \hat{u} \). As a result of having estimated the friction velocity \( u \) using our numerical results, the virtual friction velocity \( \hat{u} \) approximately agrees with the real friction velocity \( u \). Thus, hereafter, we describe \( \hat{u} \) as \( u \). The wave orbital velocity is given on the bottom wave surface. The passive scalar \( T \) is set a constant value \( T = 1 \) at the upper boundary (\( z = h \)) and \( T = 0 \) at the bottom boundary (\( z = 0 \)).

![Fig. 1. Computational configuration.](image-url)
The coordinate system \((\xi, \eta, \zeta)\) used for the computation is the general orthogonal coordinate system curving along the wave surface. The mapping from the physical space to the computational space is defined by (Benjamin, 1959)

\[
\begin{bmatrix}
\xi \\
\eta \\
\zeta
\end{bmatrix} = \begin{bmatrix}
x - ia \exp[-k(z - ix)] \\
y \\
z - a \exp[-k(z - ix)]
\end{bmatrix},
\]

where \(i\) is an imaginary unit, \(a\) and \(k\) are the amplitude and the wave number \((k = 2\pi / \lambda, \lambda:\) wave length) of the traveling waves, and the physical quantity is denoted by the real part of (4). The origin of \(\xi\) \((\xi = 0)\) is taken at the wave surface, and the value of \(\zeta\) approaches the value of \(z\) as \(\zeta\) increases. The shape of the wave surface \(z_{\text{bar}}(x)\) can be obtained by substituting \(\zeta = 0\) into equation (4), giving its approximate value to the first order of \(ak\) by

\[
z_{\text{bar}} = a \cos(kx) - a^2 k \cos^2(kx).
\]

The details of the coordinate transformation and the numerical method are described in Kihara et al. (2006). Since we want to analyze the airflow whose non-linearity is weak, the wave steepness is a constant gentle value \((ak = 0.1)\). We carry out the simulations for the 10 wave ages \(c/u\). \((c/u = 0, 2, 4, 5, 6, 8, 10, 12, 16\) and 20). The Reynolds number \(Re_c = u_h / \nu\) is set to be 150.

3. Definition of Averages

The variance whose wave number is the same as the bottom wave surface is generated in the streamwise direction \((x)\) in the airflow over a two-dimensional wave surface. Therefore, to analyze the numerical results a physical quantity \(q(\xi, \eta, \zeta, t)\) is decomposed into a phase-averaged component \(\bar{q}(\theta, \zeta)\), which is obtained by averaging \(q\) over the spanwise direction \(y\), time \(t\) and all the waves contained in the wave train, and a turbulent component \(q'(\xi, \eta, \zeta, t)\):

\[
q(\xi, \eta, \zeta, t) = \bar{q}(\theta, \zeta) + q'(\xi, \eta, \zeta, t),
\]

\[
\bar{q}(\theta, \zeta) = \frac{1}{NL_x \Delta t} \sum_{n=0}^{N-1} \iint q(\xi = \lambda\left(n + \frac{\theta}{2\pi}\right), \eta, \zeta, t) \, d\eta dt,
\]

where \(\Delta t\) is the averaging period of time, and \(N (=6\) in the present study) is the number of traveling waves existed in the wave train. The wave crest is located at \(\theta = 0\) and \(2\pi\), and the trough is at \(\theta = \pi\). To clarify the streamwise variance of \(\bar{q}(\theta, \zeta)\), the phase-averaged component \(\bar{q}(\theta, z)\) is further decomposed into an ensemble-averaged component \(\langle q \rangle(z)\), which is obtained by averaging \(q\) over \(x, y,\) and \(t\), and a wave-correlated component \(\tilde{q}(\theta, \zeta)\):

\[
\bar{q}(\theta, \zeta) = \langle q \rangle(\zeta) + \tilde{q}(\theta, \zeta),
\]

\[
\langle q \rangle(\zeta) = \frac{1}{L_x L_y \Delta t} \iint q(\xi, \eta, \zeta, t) \, d\xi \eta dt = \frac{1}{2\pi} \int_0^{2\pi} \bar{q}(\theta, \zeta) \, d\theta.
\]
4. Results and Discussions

4.1 Roughness Height for Momentum Transfer

In order to discuss the relation between the momentum transfer and the scalar transfer, we first investigate the roughness height for the momentum transfer. Investigating the dependence of the mean velocity at the logarithmic layer on the wave age, we estimate the nondimensional roughness height $z_{0m}^*(=z_{0m}g/u_*^2)$ for the momentum transfer, where $g$ is the acceleration of gravity and $z_{0m}$ is the roughness height for the momentum transfer and is related to the mean velocity at the logarithmic layer ($k\zeta > 1.5$ in the present study) as follows:

$$\langle u \rangle + c = \frac{u_*}{\kappa} \ln \left( \frac{\zeta}{z_{0m}} \right),$$  \hspace{1cm} (10)

where $\kappa$ is the von Kármán constant. The wave velocity for the deep water wave is given as $c = \sqrt{g/k}$. Substituting this relation into the definition of the nondimensional roughness height $z_{0m}^*$, $z_{0m}^*$ is rewritten as

$$z_{0m}^* = \left( \frac{c}{u_*} \right)^2 k z_{0m}.$$  \hspace{1cm} (11)

Substituting $k z_{0m}$ into (11), which is estimated by using the mean velocity $\langle u \rangle + c$ at the center of channel ($k\zeta = 2.4$), we can obtain the nondimensional roughness height. In Fig. 2, we show $z_{0m}^*$ as a function of the wave age $c/u_*$. In the present study, the Reynolds stress is very low, and the wave surface is smooth, so that the numerical conditions are significantly different from those of open ocean. Thus, the quantitative discussions will be meaningless, so they are omitted, but we just show the results of earlier works by Toba et al. (1990), Smith et al. (1992), Charnock (1955) ($z_{0m}^* = 0.012$) in the figure together.

---

![Figure 2](image-url)  

Fig. 2. Wave-age dependence of the non-dimensional roughness height for momentum transfer.  
•: the present study.
4.2 Roughness Height for Scalar Transfer

Next, in order to investigate the wave-age dependence of scalar at the logarithmic layer, we also evaluate the nondimensional roughness height for the scalar transfer $z_{0T}$ (*$z_{0T}$*), where $z_{0T}$ is the roughness height for scalar transfer and is related to the mean scalar $\langle T \rangle$ at the logarithmic layer ($k\zeta > 1.5$ in the present study) where the effect of the bottom waves was not observed:

$$\langle T \rangle = \frac{T}{K} Pr_t \ln \left( \frac{z}{z_{0T}} \right),$$

(12)

where $Pr_t$ is the turbulent Prandtl number, which is the ratio of the eddy kinematic viscosity $\nu_t$ to the eddy diffusion coefficient $D_t$, i.e., $\nu_t / D_t$. The nondimensional roughness height $z_{0T}$, as well as $z_{0m}$, is rewritten as

$$z_{0T} = \left( \frac{c}{u_*} \right)^2 k z_{0T}.$$  

(13)

In order to estimate $z_{0T}$ from (12), we need to evaluate the turbulent Prandtl number $Pr_t$. In general, a constant value is used for $Pr_t$ in the numerical simulations; $Pr_t = 0.85-1.0$ in the case of wall turbulence, and $Pr_t = 0.5-0.7$ in the case of free turbulence. In Fig. 3, we show the vertical distributions of $Pr_t$ obtained for $c/u_* = 0, 2, 6, 12$ and 20 in our DNS. Fig. 3 shows that the turbulent Prandtl number is approximately a constant value of 1.0, except the near-wall region ($\zeta < 5$). This indicates that there is the similarity of turbulent transfer between the momentum and the scalar.

![Fig. 3. Vertical distributions of the turbulent Prandtl number.](image)

Fig. 3. Vertical distributions of the turbulent Prandtl number. ○, $c/u_* = 0$; ■, $c/u_* = 2$; □, $c/u_* = 6$; ▲, $c/u_* = 12$; □, $c/u_* = 20$. 
Similarity between Momentum Transfer and Scalar Transfer

Fig. 4. Comparison between the wave-age dependence of the non-dimensional roughness height for the scalar transfer (●) and the momentum transfer (○).

The roughness height for the scalar transfer $z_{0r}$ is estimated from Eq. (12) with $Pr_i=1.0$, and we can estimate $z_{0r}^*$ from Eq. (13) using above-obtained $z_{0r}$. We show in Fig. 4 $z_{0r}^*$ along with $z_{0m}^*$. Fig. 4 shows that the same wave-age dependence is observed between $z_{0r}^*$ and $z_{0m}^*$. This means that there is a strong similarity between the momentum transfer and the scalar transfer in the airflow over the traveling waves.

5. Conclusions

Carrying out the three dimensional direct numerical simulation of the airflow over a traveling wave train moving with a constant speed, we investigated the effects of the traveling waves on the momentum transfer and the scalar transfer in the airflow. Simulations for 10 wave ages in the range of $0 \leq c/u \leq 20$. The numerical results showed that there is the similarity of turbulent transfer between the momentum and the scalar. And, the wave-age dependence of the momentum transfer was the same as that of the scalar transfer. This indicates that there is the similarity between the momentum transfer and the scalar transfer in the airflow over a traveling wave train.

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Statistics of Quasi-geostrophic Vortex Patches

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Abstract
The statistics of quasi-geostrophic point vortices and spherical vortex patches are investigated theoretically and numerically, in order to understand fundamental aspects of quasi-geostrophic turbulence. Very large numerical computations (N=2000) are performed. The most probable distributions are determined based on the maximum entropy theory. The theoretical predictions agree well with the numerical results.

Keywords: Point vortices, Spherical vortex patches, Quasi-geostrophic turbulence, Maximum entropy theory

1. Introduction

Geophysical flows are under strong influence of the buoyancy force and the Coriolis force. Due to the effects of Coriolis force and stable stratification, vertical motions are suppressed. At the lowest order of approximation, geophysical flows are considered to be two-dimensional. There have been many theoretical and numerical studies on the two-dimensional point vortex system. The statistical mechanics of point vortices was investigated by Onsager (1949), Joyce and Montgomery (1973), Kida (1975) and Lundgren and Pointin (1997). Recently, Yatsuyanagi et al. (2005) performed a very large numerical simulation of 2D point vortices (N=6724), and investigated their statistical properties. Actual geophysical flows are three-dimensional. The next order approximation of geophysical flows, is the 'quasi-geostrophic approximation', which incorporates this three-dimensionality. Under the quasi-geostrophic approximation, Miyazaki et al. (2001) developed an ellipsoidal vortex model, in which each vortex is modeled by an ellipsoidal patch of uniform potential vorticity embedded in a 'locally uniform shear field' induced by other vortices (Meacham's ellipsoidal solution). Later, Li et al. (2006) refined the ellipsoidal model, by introducing a new set of nearly canonical variables with clear geometrical meaning. Thus a natural sequence of vortex models from the point model (N degrees of freedom) through the wire-(spheroidal) model (2N degrees of freedom) to the ellipsoidal model (3N degrees of freedom) was constructed.

In this paper, we investigate the statistical properties of quasi-geostrophic point vortices and spherical vortex patches both theoretically and numerically. Numerical simulations of N-vortex system (N=2000) in an infinite fluid domain are performed. The most probable distributions are determined based on the maximum entropy theory. The theoretical predictions agree quite well with the numerical results.
2. Quasi-geostrophic Approximation

We can introduce a stream function $\psi(x, y, z)$, since the fluid motion in each vertical plane is two-dimensional:

$$u = \partial \psi / \partial y, \quad v = -\partial \psi / \partial x.$$  \hspace{1cm} (1)

The vorticity time-evolution is governed by,

$$\left( \partial / \partial t + \left( \partial \psi / \partial y \right) \left( \partial / \partial x \right) - \left( \partial \psi / \partial x \right) \left( \partial / \partial y \right) \right) \eta = 0,$$  \hspace{1cm} (2)

where $\eta$ denotes the potential vorticity:

$$\eta = -\Delta \psi = -\left( \partial^2 / \partial x^2 + \partial^2 / \partial y^2 + \partial^2 / \partial z^2 \right) \psi.$$  \hspace{1cm} (3)

In the point vortex system, we assume that the vorticity is concentrated on $N$ points ($R_i, i = 1, 2, ..., N$):

$$\eta = \sum_{i=1}^{N} \Gamma_i \delta(r - R_i).$$  \hspace{1cm} (4)

3. Statistical Mechanics of Quasi-geostrophic point vortices

3.1 Equations of Motion

The interaction energy $H_{\text{int}}$ between two point vortices located at $(X_i, Y_i, Z_i), (X_j, Y_j, Z_j)$ with strength $\Gamma_i, \Gamma_j$ is written as,

$$H_{\text{int}} = \Gamma_i \Gamma_j / (4\pi |R_i - R_j|), \quad \text{where } R_i = (X_i, Y_i, Z_i).$$  \hspace{1cm} (5)

The Hamiltonian of an $N$ point vortex system is given as a summation of the interaction energy of $N(N-1)/2$ vortex-pairs:

$$H = \sum_{(i,j)} H_{\text{int}}.$$  \hspace{1cm} (6)

We have the canonical equations of motion for the $i$-th vortex:

$$\frac{dX_i}{dt} = \sum_{(i,j)} \frac{1}{\Gamma_i} \frac{\partial H}{\partial Y_i} = \sum_{(i,j)} \frac{\Gamma_j}{4\pi} \frac{(Y_j - Y_i)}{|R_i - R_j|^3},$$

$$\frac{dY_i}{dt} = \sum_{(i,j)} \frac{1}{\Gamma_i} \frac{\partial H}{\partial X_i} = \sum_{(i,j)} \frac{\Gamma_j}{4\pi} \frac{(X_i - X_j)}{|R_i - R_j|^3}. \hspace{1cm} (7)$$

The center of vorticity $(P, Q)$ and the angular momentum $I$ are conserved besides the energy $H$ (Hamiltonian itself). We shift the coordinate origin to the vorticity center and the length scale is normalized using $I$:

$$P = 1/N \sum_{i=1}^{N} \Gamma_i X_i = 0, \quad Q = 1/N \sum_{i=1}^{N} \Gamma_i Y_i = 0, \quad I = \sum_{i=1}^{N} \Gamma_i (X_i^2 + Y_i^2) = \sum_{i=1}^{N} \Gamma_i.$$  \hspace{1cm} (8)

The time is scaled by the potential vorticity.
3.2 Statistics of Point Vortices

We trace numerically the time evolution of point vortices (\(N=2000\)) with strength \(\Gamma_{1,2,\ldots,N} = 0.5\), located randomly (and uniformly) in a cube initially. We call this case the "continuous case", below. The total energy is taken to be \(E(=H/N^2) = 0.123\), which is the most probable value of \(10^6\) ensembles produced randomly with the fixed angular momentum \(I\). We investigate the statistics of the equilibrium state, which is attained after \(t = 10 \sim 20\), by averaging the numerical data during \(t = 20 \sim 200\). (Fig.1)

The equilibrium state is axisymmetric and the probability distribution \(F(r, z)\) is a function of the radial coordinate \(r\) and the vertical coordinate \(z\). We can see, in Fig.2, that the equilibrium distribution is almost \(z\)-independent for \(|z| \leq 1.459\), whereas it becomes more concentrated near the axis as \(|z|\) increases ("end effect"). The distributions near the center region (\(|z| \leq 1.459\)) and near the lids (\(1.459 \leq |z| \leq 2.432\)) are approximately fitted by the following functions, respectively:

\[
F(r) \propto \exp(C_1r^{3.67}) \quad \text{(center)} \quad F(r) \propto \exp(C_2r^{2.87}) \quad \text{(lids)}
\]

Here, \(C_1, C_2\) are appropriate constants. The distribution of the center region is essentially that of a two-dimensional case.

In order to see the characteristics of the "end-effect" clearer, we performed additional numerical simulations, i.e., "one-layer" (\(z_{1,2,\ldots,N} = 0, N = 100, \Gamma_{1,2,\ldots,N} = 0.01, E = 0.0481\)) and "two-layers" (\(z_{1,2,\ldots,N} = -h/2, z_{N+1,\ldots,N+2,\ldots,N} = h/2, N = 1000, \Gamma_{1,2,\ldots,2N} = 0.01, E = 0.150\)). The equilibrium distribution function of the "one-layer" case is
approximately \( F(r) \propto \exp(C_3 r^{2.78}) \), whereas for the “two-layers” case it behaves like \( F(r) \propto \exp(C_4 r^{3.10}) \). The former result is similar to that of the “lid-regions” of the continuous case. The latter result shows intermediate tendency due to the interaction between two layers.

### 3.3 Maximum Entropy Theory

The equilibrium distribution is determined theoretically, based on the maximum entropy theory, which was applied to the system of two-dimensional point vortices by Kida (1975). We assume that \( \hat{N} \) vortices are placed continuously in the vertical range \( z_1 \leq z \leq z_2 \). The strength of each vortex is taken to be unity. The energy \( E \) is a parameter that determines equilibrium distribution, for the fixed angular momentum \( \hat{I} = I/\hat{N} = 1 \). The number density \( n(x,y,z,t) \) is related to the probability distribution function \( F(x,y,z,t) = n/\hat{N} \). The vertical distribution of vortices remains unchanged,

\[
P(z) = \iiint F(x,y,z) \, dx \, dy,
\]

because each vortex moves only in the horizontal plane where it is located initially.

The equilibrium distribution satisfies the following nonlinear integral equation that is derived based on the maximum entropy theory extended to the quasi-geostrophic flow:

\[
\log F + 1 + \alpha(z) + \beta(x^2 + y^2) + \gamma/4\pi \iiint (F(r)/|r-r'|) d^3r' = 0,
\]

\[
8\pi H \hat{N}^2 = \iiint \iiint (F(r_0)F(r)/|r-r'|) d^3r' = 0.
\]

Here, \( \alpha(z) \), \( \beta \) and \( \gamma \) are Lagrange’s undetermined constants, corresponding to the invariants \( P(z) \), \( I \) and \( E \), respectively. For layerwise vertical distributions, the governing equations (10) – (11) become much simpler.

### 3.3.1 One Layer

The equilibrium state is axisymmetric \( F(x,y) = F_1(r) \). We solved the above equations numerically for \( P(z) = \delta(z) \) and the result \( F_1(r) \) is in good agreement with that of the direct numerical simulation (Fig.3). As the energy \( E \) is increased, the probability distribution becomes concentrated near the origin more tightly.

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![Fig. 3. Equilibrium distribution for radial probability density function for 1-layer; ○: numerical result, lines: theoretical results.](image1)

![Fig. 4. Same as Fig. 2 but for 2-layers; ○: numerical result, line: theoretical result.](image2)
Fig. 5. Same as Fig. 2 but for 3-layers: solid line; theoretical result for $z = 0$, dotted line: theoretical result for $z = \pm 1$.

3.3.2 Two Layers

Assuming that the equilibrium state is axisymmetric and that the distribution is common for both layers ($F(x, y, z = \pm h/2) = F_2(r)$), we solved (10) ~ (11) for

$$P(z) = 1/2 \delta(z + h/2) + \delta(z - h/2).$$

The equilibrium $F_2(r)$ agrees well with that of the direct numerical simulation (Fig.4). As the vertical distance $h$ is changed from 0 to $\infty$, the equilibrium distribution changes from that of one layer ($h = 0$) through a rather flatter shape ($h = 1$) and back to that of one layer ($h = \infty$).

Similarly, we can determine the most probable distribution of the three-layered vortices ($z = 0, \pm h/2$), based on the maximum entropy theory. We see, in Fig.5, that the distribution of the center plane ($z = 0$) looks similar to that of the center region of continuous case, whereas the distribution of upper and lower planes ($z = \pm 1$) is more concentrated around the axis of symmetry, showing the end-effect for lid-regions in the continuous case.

4. Statistical Mechanics of Spherical Vortex Patches

The point vortex system is an inviscid system without energy dissipation. In order to incorporate dissipative processes, such as merging of vortices of the same sign, we introduce a spherical vortex patch instead of a point vortex. The initial radius is taken to be $r_s = 0.036$, which is about 30% of the most probable distance between two neighboring vortices. The initial strength is $N = 1$. The number of vortex patches in the following simulation is $N = 1000$.

4.1 Merger of Quasi-geostrophic Spherical Vortices

The motion of vortex centroid is governed by (7) with the Hamiltonian (6), if the vortices are far apart. The self-energy $H_{si}$ of a vortex patch of radius $r_{si}$ is given by Chandrasekhar (1987), as,

$$H_{si} = \frac{2\pi}{15}r_s^6 q^2 \int_0^\infty ds/\left(s + r^2\right)^{3/2} = \left(\frac{4\pi}{15}\right)r_s^2 q^2.$$

The total energy is the sum of the self-energy $\sum_{i=1}^N H_{si}$ and the interaction energy, but the self-energy has no contribution to the motion of vortex centroid. The total enstrophy $S$
is defined as,

\[ S = \sum_{i=1}^{N} \hat{\Gamma}_i \hat{\Gamma}_i = \left(4\pi/3\right)\left(r^3 S q\right). \]  

(13)

Direct numerical simulations of the quasi-geostrophic equation (based on the CASL-algorithm) indicate that two co-rotating vortices of radii \( r_i, r_j \) merge when two conditions are satisfied:

- horizontal distance \( a < 1.3(r_i + r_j) \)
- vertical distance \( h < r_i + r_j \).

The energy \( H \) and the enstrophy \( S \) are dissipated during merger, and a new vortex of radius \( r_k \) appears after merger (Miyazaki et al., 2002). We introduce a simple rule for such processes.

\[ F_S(S_i + S_j) = S_k \left(0 < F_S < 1\right). \]

(14)

Here, \( F_S \) is a constant decaying factor. This relation determines the radius of the new vortex, whose centroid coincides with that of the vortices before merger.

\[ R_k = \left(\hat{\Gamma}_i R_j + \hat{\Gamma}_j R_i \right) / (\hat{\Gamma}_i + \hat{\Gamma}_j). \]

(15)

### 4.2 Enstrophy and Energy Dissipation

In CASL-simulations, the energy decay rate was always smaller than the enstrophy decay rate, i.e., \( -\Delta S/S > -\Delta H/H \) (Li et al. (2006)). We adjust the enstrophy decaying factor \( F_S \) so that the same inequality holds. We find out that this occurs in a narrow range around \( F_S \approx 0.98 \). For \( F_S = 0.988 \), \( r_{5(0,1,2,\ldots,y)}(0) = 0.036 \), we observe power law decays of \( N(t) \sim t^{-0.277}, S(t) \sim t^{-0.00636}, H(t) \sim t^{-0.00407} \) during \( 3 \leq t \leq 40 \) (Fig.6):

![Fig. 6. Time evolution of E,N,S.](image)

![Fig. 7. Radial vorticity \( \omega(r) \).](image)

The merging process terminates at \( t \approx 40 \), and no power law behavior is seen after that. For \( F_s \leq 0.96 \), the energy decay rate is larger than that of the enstrophy, i.e., \( -\Delta S/S < -\Delta H/H \). When \( F_S \) is nearly unity, the energy increases, which is clearly unphysical. The choice of \( F_S \) is crucial for a physically meaningful simulation.

### 4.3 Statistics of Spherical Vortex Patches

Finally, we consider statistical properties of the vortex patches. The vorticity distribution \( \omega(r) \) becomes uniform (Fig.7), for the case of \( F_S = 0.988 \). This indicates that a uniform
vortex patch is formed due to weak dissipative processes, in contrast to the equilibrium distribution is formed in completely inviscid vortex interactions.

5. Summary

We have investigated the statistics of quasi-geostrophic point vortices and spherical vortex patches, and found the followings.

- Point vortices
  - The continuous equilibrium distribution is formed after $t = 10 \sim 20$.
  - The equilibrium distribution of the center region (vertically) is similar to that of two-dimensional point vortices, whereas the distribution near the upper and lower lids suffer from end-effect and concentrates tightly around the axis of symmetry.
  - The predictions based on the maximum entropy theory are in good agreement with the numerical equilibrium distributions for discrete layer cases. As the energy $E$ increases, the equilibrium distribution concentrates more and more to the axis of symmetry.

- Spherical vortex patches
  - For an appropriate enstrophy decay factor, i.e., $F_s = 0.988$, power law decays of $N, S, E$ are observed.
  - The vorticity distribution becomes almost uniform, indicating that weak dissipative processes produce a vortex patch of uniform potential vorticity.
  - In the time range between which the power law behavior is found, the normalized vortex volume distribution obeys an exponential law, indicating the occurrence of a self-similar decaying process.

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Properties of Perturbations Obtained by Nudged Ensemble Simulations of a Relatively High-Resolution Atmospheric General Circulation Model

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Abstract
The properties of the perturbations obtained in a preliminary numerical experiment of breeding of growing modes with a relatively high-resolution (T159L48) atmospheric general circulation model have been examined by investigating the spatial structures and energy spectrum of the perturbations. In the present breeding experiment, a control run is nudged to the analysis states provided by the Japan Meteorological Agency, and ninety-nine perturbations are bred along the control run through the process of globally orthogonalizing and rescaling every 12 hours. It is found that the perturbations have large amplitudes in the extratropics and their dominant spatial scales are slightly smaller than the synoptic scales. They seem to have been grown rapidly in the locally unstable areas induced by the development of mid-latitude depressions.

Keywords: Simulation with an atmospheric general circulation model, Breeding of growing modes, Ensemble experiment

1. Introduction

It is well known that there are three main causes of imperfectness in numerical weather forecasts. One is the error contained in the initial values, which originates in the observational and analysis errors. The second is the imperfectness of the forecast models. The third is the instability of the atmosphere, which is expressed in the words “sensitive dependence on initial conditions”, “butterfly effect”, or “chaos”. While the errors due to the above first and second causes can be reduced rather straightforwardly by establishing more effective observation network and developing more precise numerical models, there seem no straightforward ways to overcome the third because it is an intrinsic nature of the atmosphere. However a better understanding of the characteristics of the atmospheric instability should give helpful suggestions for reducing the third cause. In this study, we have investigated the properties of the perturbations obtained by nudged ensemble simulations of a relatively high-resolution atmospheric general circulation model (AGCM). Based on this preliminary investigation, we plan to perform systematic numerical experiments for a better understanding of the characteristics of the atmospheric instability.
2. Properties of Breeding of Growing Modes

The perturbations, whose properties we have investigated, are generated by applying the method of breeding of growing modes (see Toth and Kalnay, 1997) to an AGCM, called AFES (AGCM for the Earth simulator; Ohfuchi et al., 2004) with a spatial resolution of T159L48 (80-km horizontal resolution and 48 vertical levels). A control run of the breeding experiment is forced to follow the evolution of the real atmosphere from March 1st to April 10th, 2004, by replacing the states of the control run with the global analysis states (GPV data) provided by the Japan Meteorological Agency every 12 hours. Ninety-nine perturbation runs are performed by replacing the perturbations, which are the deviations from the control run, at the same time when the control run is replaced (i.e., every 12 hours). The replaced perturbations are globally orthogonalized with respect to the energy norm and rescaled to the magnitude corresponding to about 1 m/s in wind speed.

Ninety-nine Lyapunov exponents estimated from the present breeding experiment are all positive, which indicates that the number of modes growing in the AGCM is more than a hundred. A huge number of members are required for an ensemble experiment to reveal all the details of the atmospheric instability. Doubling times of the ninety-nine breeding perturbations are estimated from 1.5 to 2.5 days. The breeding perturbations have large amplitudes in the extratropics, which seem to grow in the locally unstable areas induced by the development of mid-latitude depressions (Fig. 1). The analysis of spatial energy spectrum shows that the dominant spatial scales of the breeding perturbations are smaller than the synoptic scales and tend to become smaller with the breeding mode number.

Fig. 1. 500 hPa geopotential height for the control run (contour) and perturbations of 500 hPa geopotential height for the first breeding mode (color) at 00UTC on April 10th, 2004. Contour interval is 60 m.
3. Discussions

The above results are based on a preliminary breeding experiment. The properties of the perturbations evolving in an AGCM may depend on the conditions under which the experiment is performed. For example, when the time interval of the breeding cycle is shortened, or the magnitude of the replaced perturbations is made smaller, another type of instability, such as the topographic instability and the convective instability, which might result in saturation in the present experiment, may dominate in the breeding perturbations. We will examine the structures of evolving perturbations systematically by performing numerical experiments with various conditions in order to understand the atmospheric instability in more detail.

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4. Other Research
Application of Sediment Routing Model through Forest to Agricultural Area in Kuchoro River Watershed, Hokkaido

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Abstract
In order to preserve and manage the water environment and landscape in the watershed, the sediment control from upstream to downstream is one of indispensable managements. The watershed consists of river network and various kind of land use. In this paper we propose the numerical model to predict the sediment runoff in forest and agricultural area due to rainfall runoff from the viewpoint of constructing the future sediment routing model in the whole watershed. By applying the proposed model to the Kuchoro river catchment flowing into Kushiro Mire, we explain that the stream wise variation of watershed land use has a significant influence on both the sediment yield and the concentration of suspended solids through the watershed.

Keywords: Sediment routing, Sediment yield in forest and agricultural area, Riverbank erosion

1. Introduction

Channel modification and improvement of drainage in the Kuchoro River watershed rapidly increased the sediment inflow to Kushiro Mire, and decreased its area to 70%. This is one of examples of sediment disaster. Over the past few decades, Kushiro Mire, the largest in Japan, has lost 30% of its area. Nakamura et al. (1997) reported that this loss is due to the deposition of suspended sediment from the Kuchoro River watershed owing to channel modification and improvement of drainage for the purposes of reclamation and cultivation in the watershed. The increase of farmland area and decrease of forest area has increased the sediment yield and outflow in the watershed. Most of the vast amount of sediment is rapidly transported into the mire. The sediment and nutrients promote the growth of vegetation in the flooded area, which in turn has led to the transition of the mire ecosystem. Hence, to prevent the spread of vegetation, it is urgent that we estimate the sediment yield and riverbank erosion. To estimate eutrophication, evaluate countermeasures against soil loss, and determine the effects of sedimentation on river morphology and habitat formation, we need to model sediment runoff, which is governed by rainfall runoff. In this study, we propose a fundamental outline of sediment runoff modeling within the watershed.

2. Outline of Sediment Routing Model

Sediment routing models usually consist of several process sub-models, and we explain the sub-models.
Riverbed Variation Model

Sediment routing changes the elevation of the flood plain and riverbed. Riverbed variation is caused by an imbalance of sediment inflow and outflow, and is written as follows:

\[
\Delta z = -(1 - \rho_0) \frac{\Delta t}{\Delta x} \cdot \{(q_{\text{in}})_{\text{n}} - (q_{\text{in}})_{\text{out}}\},
\]

in which \(\Delta z\) = difference in bed elevation, \(\rho_0\) = porosity of sand, \(\Delta t\) = discrete time with respect to calculation, \(\Delta x\) = unit reach length, and \((q_{\text{in}})_{\text{n}}\) and \((q_{\text{in}})_{\text{out}}\) = transport rate of bed material load. Because total load can be estimated from the bed load transport rate and the suspended load transport rate for higher bed shear stress, we used an equation proposed by Laursen (1958). Assuming that the friction law of flow follows the Manning–Stickler equation and the effective bed shear stress can be replaced by the total bed shear stress on a flat riverbed, the original version by Laursen (1958) can be modified in the following simple expression:

\[
q_{\text{in}} = \frac{7.66}{265} \sqrt{\tau_z} \left[ \frac{\tau_z}{\tau_*} - 1 \right] f_B \left( 1 + f_{BS} \right),
\]

in which \(q_{\text{in}}\) = total load, \(q_{\text{in}} = q_{\text{in}} / \{\sigma / (\rho - 1)gd^3\}^{0.5}, \tau_* = u_*/(\sigma / (\rho - 1)gd)\) = dimensionless bed shear stress, \(u_*\) = frictional velocity, \(\sigma\) and \(\rho\) = densities of sand and water, \(g\) = gravitational acceleration, and \(d\) = sand diameter. \(f_B\) indicates the ratio of bed load transport rate, and \(f_{BS}\) indicates the ratio of suspended load to bed load. \(f_B\) and \(f_{BS}\) can be approximated on the basis of the empirical graph obtained by Laursen (1958):

\[
f_B = 10.67 \left( \frac{u_*}{u_0} \right)^{0.25}, \quad 1 \leq u_*/u_0 < 20;
\]

\[
f_{BS} = 0.383 \left( \frac{u_*}{u_0} \right)^2, \quad u_*/u_0 > 20;
\]

\[
f_{BS} = 900 \left[ 1 - \exp \left( - \frac{1}{1500} \left( \frac{u_*}{u_0} \right)^2 \right) \right],
\]

in which \(u_0\) = settling velocity of sand, and \(u_*/u_0\) is usually calculated with the formula of Rubey (1933):

\[
\frac{u_*}{u_0} = \frac{\sqrt{\tau_z}}{\sqrt{\frac{2}{3} + \frac{36}{d}} - \frac{36}{d}},
\]

\[
d_* = (\sigma / (\rho - 1)gd^3)/\nu^2, \quad \nu = \text{kinematic viscosity of water. For higher bed shear stress} \ (\tau_* \to \infty), \ q_{\text{in}}\) is approximately proportional to \(\tau_*^2\).

Transport of Wash Load

Though Einstein et al. (1940) pointed out that wash load cannot be evaluated on the basis of local flow conditions and properties of bed materials, we used the following equation proposed by Egashira and Matsuki (2000) to consider the intense effect of deposition of fine sediment on geomorphic changes to the riverbanks and flood plain:

\[
\frac{\partial c_s}{\partial t} = \frac{Q_{\text{in}} + (c_s Q)_{\text{n}} - (c_s Q)_{\text{out}}}{ALR},
\]

in which \(c_s\) = volumetric concentration of wash load, \(t\) = time, \(Q\) = flow discharge, \(A\) =
cross-sectional area of flow, \( L_R \) = length of unit reach (defined as the channel between upstream and downstream conjunctions), \( Q_{SS} \) = lateral inflow rate of sediment yield per unit length in catchments slopes and riverbanks, and subscripts in and out indicate the inflow at the upstream end and the outflow at the downstream end of a unit reach. Since the function of \( f_{SS} \) was empirically determined in a flume experiment by means of fine sand with a diameter of 0.01 to 0.2 mm, we assumed that this formula could express the transport rate of the wash load.

**Rainfall Runoff from Catchments Slopes (Stanford Watershed Model)**

We selected the Stanford watershed model, developed by Crawford and Linseley (1966), as the most appropriate rainfall runoff model. It is a conceptual distributed hydrologic model that can take into account the spatial distribution of land use, soil structure, and soil texture and is faithful to the concept of material transport by water. Because it accounts for long-term rainfall runoff, it can evaluate the soil moisture in the surface layer, which determines the precision of short-term runoff. It uses a kinematic wave runoff model under brief and unsteady precipitation that can physically estimate the surface runoff that drives sediment runoff.

**Sediment Yield from Catchments Slopes due to Rainfall Runoff**

Soil erosion equation caused by rainfall runoff from agricultural slopes can be principally written as follows:

\[
q_B = \Xi \cdot \Pi \cdot \Gamma \cdot q_{B0},
\]

(9)

in which where \( q_B \) = computed transport rate (unit volume per unit width and time), \( \Xi \) = vegetation cover factor, \( \Pi \) = cropping and management factor, \( \Gamma \) = support practice factor, and \( q_{B0} \) = transport rate on bare land without support and management.

In this study we used the following equation proposed by authors:

\[
q_{B0}^* = A_0 \left\{ qI_* - (qI_*)_c \right\}^m,
\]

(10)

in which \( q_{B0}^* = q_{B0} / \{(\sigma/\rho - 1)gd^3\}^{0.5} \), \( q_* = q / \{(\sigma/\rho - 1)gd^3\}^{0.5} \), \( I_* = I / (\sigma/\rho - 1) \), \( d \) = representative diameter, \( I \) = slope gradient, \( q \) = unit discharge of surface flow at the end of the slope, and \( A_0, m, \) and \( (qI_*)_c \) = empirical non-dimensional critical stream power. In Fig. 1, the empirical parameters are \( A_0 = 1.25, m = 5/3, \) and \( (qI_*)_c = 0.002 \).

Because the Stanford watershed model uses the kinematic wave runoff model, the calculated surface flow discharge can be directly substituted into Eq. (10). Because this equation is based on results under both artificial and natural rainfall, it gives the transport rate caused by raindrop impact and surface flow. Hence, in this study we employed the above simple expression.

**Riverbank Erosion Model due to River Flow**

Riverbank erosion is a source of sediment. The quantity of fine sediment \( q_{WS} \) supplied to the river by bank erosion is the product of the transport rate along the side slope \( q_{Bp} \) and the rate of fine sediment \( Q_{p} \) (\( q_{WS} = q_T \cdot q_{BP} \)). The distribution of \( q_{BP} \) along the wetted perimeter is expressed as follows:

\[
q_{BP} = q_B \tan \phi_s,
\]

(11)

in which \( q_B \) = transport rate in the center of the channel and \( \phi_s \) = angle of moving sand particles. Without the transverse component of flow (\( \phi_s \)), it is written as follows (Nakagawa et al., 1986):
\[ \tan \phi_f = -\frac{\Omega}{\sqrt{r_s}} \cdot \frac{\partial z_b}{\partial y}, \]

in which \( \Omega = \{(2A_3 / (C_D A_2))^0.5 \), \( z_b \) = elevation of side bank, and \( y = \) transverse axis. For simplicity, we have used the value of \( \phi_f \) where the average bed shear stress occurs along the wetted perimeter.

Determination of the shape of the cross section and the expression for bed shear stress gives the average bed shear stress along the wetted perimeter of the side bank:

\[ \bar{\tau}_b = \int_0^{B_s} \tau_b(y)dy / \int_0^{B_s} \sqrt{1 + \left(\frac{\partial z_b}{\partial y}\right)^2} dy, \]

in which \( \tau_b \) = average bed shear stress, \( \tau_b(y) = \) distribution of bed shear stress, \( z_b(y) = \) elevation of side bank, and \( B_s = \) width of one side bank. In this study we assumed the shape of the side bank proposed by Ikeda (1981) and the distribution of bed shear stress as follows:

\[ z_b = H \cdot \exp(-y / \Delta), \]

\[ \tau_b = \tau_{b0} \cdot \frac{\partial z_b}{\partial y} \cdot \left(1 - \frac{y}{H}\right), \]

in which \( H = \) flow depth in the center of the channel, \( \Delta = \) shape factor, and \( \tau_{b0} = \) bed shear stress in the center of the channel.

Supposing that the angle of repose of sand at the water's edge is 45°, \( H / \Delta = 1.0 \), and \( B_S = 4\Delta \), Eq. (13) gives the average bed shear stress as follows:

\[ \bar{\tau}_b = 0.67 \tau_{b0}, \]

Given an appropriate equation for bed load discharge in the center of channel, the erosion rate along the side bank \( q_{bp} \) can be calculated. However, mountain streams, which supply a great amount of wash load, usually form a protective armor coat, and hence bank erosion is active after armor coat is destroyed. For this reason the actual transport rate \( q_{bp} \) caused by side bank erosion should be written as follows (Kanayashiki et al., 1980):

\[ \bar{q}_{bp} = 0 \quad ; \quad h \leq h_d \]

\[ \bar{q}_{bp} = (h - h_d) / h \cdot \bar{q}_{bp0} \quad ; \quad h > h_d \]

in which \( q_{bp0} \) = transport rate without reference to the destruction of the coating and \( h_d \) = flow depth at the destruction of the armor coat. \( h_d \) can be estimated by means of the critical bed shear stress for mixed sands proposed by Egiazaroff (1965):

\[ h = \left\{ \frac{\ln 19}{\ln(19d_{so} / d_m)} \right\}^2 \tau_{cm} \left(\frac{\rho - 1}{\rho}\right)d_{so}, \]

in which \( d_{so} = \) sand diameter corresponding to 90% of accumulated sand, \( d_m = \) mean diameter, \( \tau_{cm} = \) non-dimensional critical-bed shear stress for mean diameter (\( \tau_{cm} = 0.05 \)), and \( i_b = \) river gradient.

**Channel Hydraulic Geometries (Empirical Expression)**

Detailed information is often lacking on channel hydraulic geometries in sediment routing. Under such a condition it is necessary to infer the channel hydraulic geometries of channel width \( B \), riverside width \( B_s \), central flow depth \( H \), and median diameter \( d_{so} \) from the river gradient \( i_b \), flow discharge \( Q \), and fragments of information on channel hydraulic geometries. We used the following empirical equations proposed by Fujita (1981):
\[ B = 0.523d_{50}^{-12/32}Q^{9/16}, \]  
\[ H = 0.239d_{50}^{1/16}Q^{1/8}, \]  
\[ R_k = 4H, \]

in which we assumed that \( Q = (A_{W}/A_{W})Q_{\text{max}} \), where \( A_{W} \) = area of sub-watersheds, \( A_{W} \) = area of whole watershed, and \( Q_{\text{max}} \) = annual maximum flow discharge at the endpoint of the whole watershed or river network.

The median diameter \( d_{50} \) of sands in rivers in equilibrium is roughly estimated as \( d_{50} = (d_{50})_0 \cdot (i_{b}/i_{b0})^{2/3} \), when the median diameter \( (d_{50})_0 \) and river gradient \( i_{b0} \) at some point are known.

No method has been proposed to estimate the geometrical standard deviation \( \sigma_g \) of mixed sands on the basis of information on channel geometries. When the geometrical standard deviation \( \sigma_g \) of mixed sands is known, \( d_{50} \), which determines flow depth \( h_{d} \) when the protective coating is destroyed, is estimated for a log-normal sand diameter distribution as:

\[ \frac{d_{50}}{d_{m}} = \frac{d_{50}}{\text{exp}(\ln d_{50} + \ln^2(\sigma_g / 2))} \cdot \sigma_g^{1.64}. \]

**Flood Routing Model in River Network**

Because runoff flows into rivers, the flow discharge and sediment transport rates depend on how precisely the river flow is described. It is better to use a dynamic wave model for flood flow, but it is difficult to solve its equations for network structures. For this reason we used a diffusion wave model:

\[ \frac{\partial Q}{\partial t} + w \left( \frac{\partial Q}{\partial x} - q_0 \right) = \mu \frac{\partial^2 Q}{\partial x^2}, \]

\[ w = \frac{\partial Q}{\partial A} \]

\[ \mu = Q/(2B_{ik}) \]

in which \( w \) = propagation velocity, \( \mu \) = coefficient of diffusion term, \( t \) = time, \( Q \) = flow discharge, \( x \) = downward distance, \( q_0 \) = lateral inflow rate per unit length of river, \( A \) = cross-sectional area of flow, \( B \) = river width, and \( i_{b} \) = river gradient.

**Results and Discussion**

To examine the accuracy of the proposed total model and respective sub-models, we applied it to the Kuchoro River watershed. Fig. 1 and 2 shows the outline of watershed and longitudinal changes of river gradient before and after channel modification.

**Verification of Rainfall Runoff Model**

Pseudo-channel networks were generated on the basis of the grid-based 250-m Digital Elevation Model published by the Geophysical Survey Institute, Japan, and we defined unit stream sections between the nodal points of the confluence and divergence of rivers. Attributes that characterize runoff, including land use, soil texture, slope gradient, and area, were categorized, processed, and over-layered by means of geographic information system, ARC/INFO. Daily precipitation data came from the Meteorological Agency of Japan and the Hokkaido Development Council. Flow discharge records for 1990 through 1993 were used. Fig. 3 shows the validity of the Stanford watershed model as a runoff model and of the diffusion wave model as a channel network model. It indicates good agreement between the observed and calculated discharge at the Shimo Kuchoro measurement station.
Fig. 1. Outline of Kuchoro River watershed.

Fig. 2. Longitudinal changes of river gradient before and after channel modification.

Verification of Sediment Runoff Model

Nakamura et al. (1997) measured the flow discharge and concentration of the suspended load and wash load at four measuring stations St.1 ~ St.4 as shown in Fig. 2, in Kuchoro river, from 27 to 30 September 1995. They estimated the contribution of fine sediment to be 350 t from the upper watershed (upstream from St.1), 430 t from the middle watershed (from St.1 to St.2), and 680 t from the lower watershed (from St.3 to St.4). On the basis of a sediment budget for steep bank erosion, they also estimated the sediment yield from the watershed slope except the bare bank as 182 t (upstream from St.1), and 360 t (from St.1 to St.2) respectively.
Fig. 3. Comparison between the measured flow discharge and the calculated one by Stanford watershed model.

Fig. 4. Comparison between the measured concentration of wash load and the calculated concentration at St.1.

Fig. 4 compares the observed and calculated concentrations at St.1. The calculated concentration (based on the wash load model) increased sharply and decreased rapidly, which means that riverbank erosion occurs when the protective coating is destroyed. This rapid decrease is very different from the observed slow decrease, because of the difficulty of estimating the possible erosion reach and the effects of vegetation cover, cropping management, and protection works. Fig. 5 presents a simulated sediment budget in the Kuchoro watershed for the same dates. The calculated fine sediment yields were 566 t from the upper region, 240 t from the middle region, and 377 t from the lower region. We judge that the degree of precision of the models of sediment yield caused the discrepancies between the calculated values and the values measured by Nakamura et al. (1997).
3. Conclusion

The control of sediment throughout a watershed is indispensable to managing water environments. Modeling of a watershed consisting of a river network and various land uses on slopes requires an appropriate sediment routing model rather than a universal model. In this paper we propose a mathematical model to predict the sediment budget in an agricultural and forest watershed. By applying the model to the Kuchoro watershed, we explain that the variation in watershed land uses has a significant effect on sediment yields and riverbank erosion.

References


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Development and Applications of an Urban Meteorological Numerical Model in Cartesian Coordinate (2)

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Abstract
We have developed a new urban street atmospheric numerical model in Cartesian coordinate which is expected to treat any complex object (buildings, etc.) explicitly in urban boundary with a finer resolution. In this work, the numerical simulation has been conducted by Large Eddy Simulation (LES) modeling on the thermal and turbulent structures of the flow over a low-rise building residential area in a modeled urban canyon.

Keywords: Urban canopy flow, LES modeling, Cartesian coordinate, Complex object, Blocking-off method

1. Introduction

A high-resolution local atmospheric numerical model has been developed in the Cartesian coordinate, and it is expected to suitably treat the steep topography and complex objects in urban city with a finer resolution. In the model, finite volume method (FVM) in conjunction with the Semi-Implicit Method for Pressure-Linked Equation Revised (SIMPLER) algorithms is used for calculations of the unsteady, three-dimensional, compressible flows on a staggered grid. Abandoning the customary terrain-following normalization, we choose the Cartesian coordinate and the blocking-off method is introduced to handle all of the steep topography and complex objects (e.g., buildings in urban city). So far, the numerical model has been run on calculating flows over cubes, steep mountains by Direct Numerical Simulation (DNS), and turbulent flow in actual urban city by LES, respectively.

In the present work, we implement and apply the LES numerical model for modeling the stratified flow within and above a modeled urban canopy. When thinking of the microclimate or the air quality in an urban city, the airflow structure and turbulent exchange in the urban boundary layer should be pre-determined since they control the wind, temperature and pollution concentrations. The urban canopy is considered as a typical morphological unit in the urban boundary layer, and therefore much research has been focused on it in terms of understanding its dynamics and energy balance. The aims of this work are to simulate the urban canopy flow, and to examine the thermal flow structure and formation mechanics.

2. Numerical Model

3. Result

The numerical model has been run on LES modeling of an unstable stratified flow within and above a modeled urban canopy. Fig. 1 shows the geometry in the simulation. In Fig. 2, the temperature and wind velocity near the surface are shown, and characteristics of the thermal flow within the urban canopy are well simulated.

![Fig. 1. Geometry in simulation.](image)

![Fig. 2. Temperature and wind vector within the urban canopy layer.](image)
References


Publications and Presentations

Conference Reports:

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

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HEC Group, NEC Corporation

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. Fig. 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:
- 512GB 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.

Fig. 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 FY2004.

<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>1node</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>1node</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</td>
<td>5760H</td>
<td>60GB</td>
<td>1</td>
<td>2nodes</td>
</tr>
</tbody>
</table>

2) Improvement of System Use
   Support towards efficient use is mainly offered by the policy of the following i) and ii).

i) Use of ASL/SX
   With the evolution of the computing capabilities of supercomputers, there are growing and diversified demands for numerical modeling of natural phenomena and engineering problem. Fast and robust numerical algorithms are essential as a basis for scientific applications. NEC has been developing the ASL/SX Series mathematical libraries to facilitate the state-of-art numerical algorithms in building scientific and engineering application programs.

   NEC provides the ASL/SX as well as the statistical library ASLSTAT/SX and C-language interface library ASLCINT/SX as proprietary scientific libraries on the SX-6. In the explanation below, these libraries together will be called the ASL/SX Series. The ASL/SX Series covers a large variety of functions so that they can be utilized in many research and development fields. Currently, approximately 1,600 functions are provided with the ASL/SX Series (R18.0), including around 1,300 numerical functions and 300 statistical functions.

   The ASL/SX Series has been intensively vectorized and parallelized so that it can fully exploit the capabilities of the SX Series. Fig. 2 shows the parallel performances of the major parallelized subroutines of ASL/SX. ASL/SX achieved an excellent performance of around 60GFLOPS (more than 90% peak performance) for a matrix-matrix multiplication on a single-node SX-6 system consisting of eight processors, while it realized 40GFLOPS (or around 60% peak performance) for the Fast Fourier Transform. Furthermore, the performance increases almost linearly with the increasing number of processors from one to eight, thus demonstrating a superior performance scalability of the ASL/SX.
Fig. 2. Parallel performance of ASL/SX. Target machine: SX-6/8A (Maximum performance: 64GFLOPS=8CPU*8GFLOPS/CPU). The left axis – bars: performance measured in GFLOPS. The right axis – lines: relative ratio to the theoretical peak of SX-6.

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>
<thead>
<tr>
<th>Model Name</th>
<th>The contents of correction</th>
<th>Improvement ratio</th>
<th>Execution time</th>
</tr>
</thead>
<tbody>
<tr>
<td>CCSR/NIES AGCM</td>
<td>Parallelizing by sauce correction</td>
<td>5.0</td>
<td>1999.09</td>
</tr>
<tr>
<td>agem5.5p</td>
<td>Tuning to the routine out of which the performance has not come</td>
<td>1.3</td>
<td>2000.03</td>
</tr>
<tr>
<td>flood2D</td>
<td>Vector tuning Parallelizing tuning</td>
<td>49.0</td>
<td>2000.07</td>
</tr>
<tr>
<td>RAMS Ver4.3</td>
<td>Sauce correction Optimization by the compile option</td>
<td>4.0</td>
<td>2001.09</td>
</tr>
<tr>
<td>agem5.4.02</td>
<td>Parallelizing by sauce correction</td>
<td>2.0</td>
<td>2001.11</td>
</tr>
<tr>
<td>FldDynaNR_Fortran</td>
<td>Use of mathematical library: ASL Sauce correction</td>
<td>40.0</td>
<td>2002.11</td>
</tr>
<tr>
<td>MJ98-CTM</td>
<td>Parallelizing tuning</td>
<td>2.0</td>
<td>2004.04</td>
</tr>
<tr>
<td>agem5.4g.p1</td>
<td>Parallelizing by sauce correction</td>
<td>3.61</td>
<td>2005.04</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
   Fig. 3 shows supercomputer’s operation results from Apr. 2005 to Mar. 2006.

![Graph showing job count and CPU time from April 2005 to March 2006.]

Fig. 3. Supercomputer’s operation results.

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### Research Programs in FY2005

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<td>Climate change experiments with a high-resolution CGCM (NIES, The University of Tokyo, Frontier Research System for Global Change)</td>
<td>25-31</td>
</tr>
<tr>
<td>A Numerical experiments for the prediction of the future ozone layer (NIES, The University of Tokyo)</td>
<td>3-9</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>(2) Other Researches</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Development of Ozone Chemical Prediction Model (The University of Tokyo, NIES)</td>
<td>3-9</td>
</tr>
<tr>
<td>Studies on atmospheric environment with regional atmospheric models (NIES, The University of Tokyo, Nagoya University, Kyushu University, National Research Institute for Earth Science and Disaster Prevention)</td>
<td>11-17</td>
</tr>
<tr>
<td>Research on the analysis of ozone long term variations in mid-latitude and the factors affecting them (Meteorological Research Institute)</td>
<td>19-24</td>
</tr>
<tr>
<td>The high-resolution numerical model of heat island phenomena (NIES, Building Research Institute)</td>
<td>35-40</td>
</tr>
<tr>
<td>Mass transfer mechanism at and below the air-water interface in the surface ocean and the effects of droplet and swell on the mass transfer at the air-sea interface (Kyoto University)</td>
<td>41-48</td>
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<tr>
<td>Application of the transport model for inverse modeling studies of the regional and global budgets of CO₂ (Frontier Research System for Global Change, NIES)</td>
<td>49-54</td>
</tr>
<tr>
<td>Direct calculation of the interaction between cumulus convection and large-scale motions (Kyushu University, Hokkaido University)</td>
<td>55-60</td>
</tr>
<tr>
<td>Numerical simulation of the stratified/rotating turbulence and geostrophic vortices (Kyoto University)</td>
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<td>Development of atmospheric general circulation model for terrestrial planets and related fundamental experiments on the atmospheric structures (Hokkaido University)</td>
<td>69-76</td>
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<td>Atmospheric motion and air quality in East Asia (Acid Deposition and Oxidant Research Center, NIES)</td>
<td>77-83</td>
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<tr>
<td>Quasi-Geostrophic vortex motions and scalar transport (The University of Electro-Communications, Kyoto University)</td>
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<td>Ensemble simulations with huge members using an atmospheric general circulation model (Chiba Institute of Science, The Earth Simulator Center)</td>
<td>93-95</td>
</tr>
<tr>
<td>International cooperative research on the management of watershed environment (NIES)</td>
<td>99-106</td>
</tr>
<tr>
<td>Development of non-hydrostatic numerical model for the geophysical fluid dynamics of the global and regional atmosphere (Tohoku University)</td>
<td>107-109</td>
</tr>
</tbody>
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Appendices

Outlines of the Research Activity Reports (in Japanese)

Program of the 14th Supercomputer Workshop
Tsukuba, October 30, 2006
Outlines of the Research Activity Reports
(in Japanese)
CCSR/NIES T42 化学気候モデルを用いたオゾン層の将来予測実験

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1. 背景と目的
化学気候モデル（CCM）は、オゾン層の将来にわたる変動要因を明らかにするのに有用なモデルである。化学気候モデルには、大気中の力学過程、放射過程、化学過程が含まれ、また、それぞれの過程間の相互作用も加え含まれている。南極オゾンホールのような、非線形過程が働く極端な状態に高度に依存した現象の理解のために、化学気候モデルのような化学過程を含めた3次元大循環モデルを用いた研究が必要とされる。

本研究では、Nagashima et al. (2002) で開発した水平解像度 T21 (5.6° × 5.6°) 化学気候モデルをさらに発展させ、WCRP (World Climate Research Programme) 傘下の Stratospheric Processes and their Role in Climate (SPARC、成層圏プロセスとその気候における役割研究計画) の下のプロジェクト、CCMVal (Chemical Climate Model Validation、化学気候モデル評価) で提唱されているハロゲン濃度の将来シナリオに従ってオゾン層将来予測実験を行ったので、その結果について報告する。

2. モデル・数値実験設定
本研究で使用した化学気候モデルは、Nagashima et al. (2002) で使用した化学気候モデルを以下の点で改善したものである。

図 1 計算に使われた温室効果ガスとCCl4の地表面濃度の経年変化。図示の都合上、CO2とN2Oの濃度は10倍に拡大されていることに注意。

(1) 水平解像度： T21 (2.8° × 2.8°) からT42 (5.6° × 5.6°) へ増加。
(2) Hines (1997) による非地形性重力波ドラッグの導入。
(3) 臭素化合物による化学反応の導入。
(4) 極成層雲に、STS (Supercooled Ternary Solution、H2O/H2SO4/HNO3の混合液滴) を追加、同時にこの粒子に関わる不均一反応も追加。
（5）これまでの極成層雲の粒径と落下速度を固定したスキームから、極成層雲の個数密度を与えてその凝縮量から粒径を計算し、それをもとに落下速度を計算するスキームに変更。
（6）Schumann-Runge 帯（177.5 nm-202.5 nm）による物質の光解離を導入。
（7）太陽放射に対する大気の球面効果を導入。
計算は、1975年1月1日から2051年1月1日まで行った。最初の5年間は基準年とみなし、1980年1月1日以降の計算結果を解析した。CO₂、CH₄、N₂Oの将来シナリオはIPCC（2000）のA1Bシナリオを、オゾン破壊物質の将来シナリオはWMO（2003）のAbシナリオを使った（図1）。この計算は、CCMVal 推奨のREF2計算に相当するものである。この計算では、QBO、太陽11年周期、火山爆発による成層圏エアロゾルの増加の影響は除外されている。海面水温は、CCSR/NIES/FRCGCの大気海洋結合モデルのアウトプットを使った。

3. 成果
まず、前節で述べた化学気候モデルの改良によって、南極オゾン全量の10月極小が再現できるようになった。Nagashima et al. (2002)のモデルでは、11月にオゾン全量の極小が起こっており、その理由として、南極渦が安定していると、南極上空の塩素濃度が若干低かったことにによるオゾン破壊速度の過小評価が考えられる。
図2と図3に、40°S以南のオゾン全量から、その年のオゾンホールの面積の最大値とオゾン全量の極小値を求め、その経年変化をプロットした結果を示す。オゾンホールの面積は、オゾン全量が220 D.U.（ドブソンユニット）以下を示す面積で表される。

図2 オゾンホール面積（40°S以南の220 D.U.以下）の経年変化。黒がTOMS（観測値）、赤が化学気候シミュレーションによる結果。
図3 40°S以南のオゾン全量の経年変化。黒がTOMS（観測値）、赤が化学気候モデルによる結果。

オゾンホールは1980年頃から出現し、1980年代に急速に拡大し、1990年代に入ってその速度が徐々に落ち、2000年頃にピークを迎ええた様子がよく再現されている。将来にわたって計算を延長したモデルの将来予測結果は、2020年を過ぎた頃からオゾンホールの縮小傾向が見られ始め、今後50年ほどまではオゾンホールは解消するという結果になった。ただし、図からも明らかのように、大気の力学的な年々変動のため、2050年でオゾンホールが解消したという結果にはなっていない。

4. おりに
この報告ではpreliminaryな結果を示したが、今後この将来予測結果について詳しい解析を進めることで、特に、大気の力学場との関連において解析を行う。また、QBO、太陽11年周期、火山爆発による成層圏エアロゾルの増加の影響を考慮したREF1実験の結果に関する解析も予定している。
二酸化炭素増加時の温暖化に対する動態植生の影響

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東京大学気候システム研究センター

1. はじめに
人為起源の大気 CO₂濃度の上昇による将来の温暖化が予測されている。この CO₂増加と温暖化により陸上植生分布の変化、光合成・炭素分解などの炭素収支の変化を通じて大気 CO₂と温暖化に対してフィードバックが起き得ることが大気循環モデルを用いた炭素循環実験を行った先行研究により示唆されている。この陸上生態系のフィードバックの強さは先行研究で不確定性が大きいが、それは炭素循環モデルに採用した大気・海洋・生態系等の各モデル間の差異の積み重ねに起因する。これらの炭素循環温暖化実験で採用された陸上生態系モデルは植生固定のものと気候変化に伴う植生分布の変化を導入したものがあるが、植生変化による最終的なフィードバックに対する寄与は分離した議論がなされていない。

2. モデルと実験設定
本研究で大気循環モデル (AGCM) と動態植生モデル (DGVM) を結合し、植生分布の動的な変化を大気に対して反映させることを可能とした。この結合モデルを用いて大気 CO₂濃度を 285 ppm・570 ppm・1141 ppmの3通りの平衡状態実験を行った。また、比較のために従来の固定植生を採用した AGCMを用いて同様の3通りの暖化状態実験を行った。その結果を用いて植生分布を考慮した温暖化と固定植生を仮定した温暖化を比較した。これらの実験では海面温度と海水貯留は混合層モデルを用いて予報し、また、動態植生モデルを用いて各実験の出力結果から平衡状態の陸上炭素貯留量を診断し、起き得る大気 CO₂濃度に対するフィードバックを議論した。

3. 結果
動態植生の導入により、大気 CO₂濃度の増加に起因する温暖化に対して全球平均で +10%、陸上平均で +20% 増幅された（図1）。大気 CO₂濃度実験では温暖化と CO₂濃度変化によって全球的に光合成活動が活発化し、植生分布が増加している。この温暖化増幅は植生分布の変化に伴う地表面アルベドの低下に起因すると考えられる。また、温暖化時の陸上炭素貯留量は動態植生の導入により減少した。これは主として北半球中緯度における土壌炭素の分解が温暖化の増幅により促進されたためである。陸上炭素の減少を大気 CO₂濃度に換算すると約 +18% の増加となり、このフィードバックを導入した場合にはさらなる温暖化の増幅が起きることが示唆される。

図 1 285 ppm（産業革命前）実験に対する 1141 ppm（4倍 CO₂）実験の温暖化（年平均気温の差）に対する動態植生導入による寄与。
太陽11年周期変動に対する成層圈の力学的・化学的応答の化学-気候モデルによるシミュレーション

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1. はじめに
太陽11年周期変動によって紫外線強度はおよそ十数パーセント変化するが、観測値や近赤外域での変化は非常に小さく、イラディアンスの直接の変化を通じて成層圈全体や対流層に与える影響は非常にお小さい。太陽11年周期変動の成層圈下部や対流層への影響のプロセスは力学を通すと考えられているが、その過程の詳細な判読は現在は難しいわけではない。なぜなら、その変化やオゾンのシグナルは単純な構造ではなく、モデルで現実化されているわけではないからである。本研究では1980年から2004年までの過去25年の全てのフォーシングを用いて成層圈再現実験を行い、その結果を太陽11年周期変動に着目して解析したもののである。

2. モデルと解析手法
気象研究所の化学-気候モデルT42L68を使った。水平分解能は約300km、成層圈での鉛直分解能は500mである。瞬間水温（SST）、温度効果気体、フロンについては観測されたSST、海氷分布を使い、CO₂、CH₄、N₂Oとハロゲン類の濃度を地表面において全球規模に時間（日）の関数で与えた。火山エーロゾルも観測値から見積もられた表面積、消散係数、有効半径を時間の関数（日にち）で与えた。この期間の主な火山噴火はエルチョンとピナツボである。太陽紫外線変化は11年周期に伴う紫外線強度を1nmの分解能で与えた。赤道成層圏標準二年振動（QBO）はモデル自身がQBOを再現しているので特別な措置は行わなかった。

3. 結果
モデルは観測結果と定性的に一致しており、太陽活動の最大時には昇温域が上部成層圏と下部成層圏の2カ所で現われている（図1参照）。上部成層圏のシグナルは紫外線強度の増大による直接の効果であり、下部成層圏のシグナルは力学によるもので間接の効果である。高さは異なるが、オゾンにも上部成層圏と下部成層圏で濃度増大のシグナルがモデルと観測の両方に見えており、温度変化とコンセンサシントな応答である（図略）。

図1　中低緯度（60S-60N）における帯状平均温度の太陽11年周期成分の緯度－高度断面。モデル（左）、ERA40の解析（右）。等温線間隔はモデルが0.1K、解析が0.2Kである。影域は95%有意水準を示す。
20世紀前半の気温変動に対する太陽活動変動と火山活動の影響

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1. はじめに
20世紀後半に観測された地上気温上昇に関して、多くの研究が人為起源外部強制力に起因することを指摘している(Intergovernmental Panel on Climate Change, 2001)。一方、20世紀前半に観測された地上気温の上昇に関して、我々のこれまでの研究は、自然起源外部強制力の寄与が大きいことを示した(Nozawa et al., 2005)。ここではさらに進んで、自然起源外部強制力を“太陽活動の変動によるもの”と“火山噴火に伴うもの”の2種類に分けて、それぞれの相対的重要性を議論する。

2. 20世紀気候再現実験
用いたモデルは中解像CCSR/NIES/FRCGC 大気海洋結合モデル(MIROC 3.2; K-1 model developers, 2004)である。このモデルに過去の気候変化要因外部強制力を与えて、1850年から2000年まで積分する。外部強制力の与え方を変えて、下記の5種類の実験を行った。
- SOLR: 太陽からの短波入射の変動を与える。
- VLCN: 火山噴火に伴う成層圏エアロゾルの変動を与える。
- NTRL: 太陽と火山噴火の両方の強制力変動を与える。
- ANTH: 人為起源の外部強制力を与える。
- FULL: 自然起源強制力と共に人為起源の外部強制力を与える。

各実験とも初期値の異なる4アンサンプルを計算し、そのアンサンプル平均を解析に用いる。比較する観測データとしてはJones and Moberg (2003)の地上気温データを使う。

図1は地球平均地上気温偏差(1881-1910 平均からの差)の時系列である。観測、NTRL、SOLR、VLCN実験とも20世紀前半に気温上昇が見られる。

図1 全球平均地上気温偏差(1881-1910 平均からの差)の時系列。黒線は観測。赤線はモデルの4アンサンプル平均。ピンクの陰影は4アンサンプルの最大・最小を示す。
3. 多重回帰分析

太陽活動変動と火山噴火が20世紀前半（1900-1949）の地上気温変化に及ぼした影響の有効性を評価するために、ここではoptimal fingerprinting（Allen and Stott, 2003）と呼ばれる重回帰分析手法を用いる。観測された気候変動パターン（地上気温偏差の時空間ベクトル）をyとし、ANTH、SOLR、VOLC実験のシグナルをそれぞれ$x_1, x_2, x_3$として、下記の多重回帰式を解く。

$$y = (x_1 - v_1) \beta_1 + (x_2 - v_2) \beta_2 + (x_3 - v_3) \beta_3 + v_0.$$ (1)

ここで、$v_0$は観測に含まれるノイズであり、$v_1, v_2, v_3$はアンサンプル数が有限であるために生じるノイズである。求めた際の回帰係数$\beta_1, \beta_2, \beta_3$は、スケーリングファクターと呼ばれる。スケーリングファクターの不確実性の大きさは、外部強制力を産業革命前の条件に固定した長期（1300年）のコントロール実験を用いて推定する。

図2(a)に、求められたスケーリングファクターを示す。SOLR、VLCNとも不確実性の幅に0を含まず、観測データの中にSOLR、VLCN強制力に対する有意な応答が見られることを示している。またスケーリングファクターが1に近いことから、外部強制力に対する現実での応答を、モデルが正しい振幅で再現していることがわかる。一方、この期間におけるANTHの応答が小さいため、十分なシグナル・ノイズ比を得られない。そのためANTHのスケーリングファクターは不確実性の幅が大きく、ANTHの影響を評価することができない。図2(b)に全球平均トレンドを示す。SOLR、VLCNのトレンドは、観測の約半分で、両者の和で20世紀前半に観測された地上気温変動を説明できる。

![スケーリングファクターとトレンド](image)

図2 スケーリングファクターとトレンド。(a) ANTH（緑）、SOLR（橙色）、VLCN（紫）のスケーリングファクター。エラーバーは5-95%の不確実性の幅を示す。(b) ANTH（緑）、SOLR（橙）、VLCN（紫）、全体モデル応答（黄色）、観測（黒）の全球平均トレンド。

参考文献


規則的に配列された建物群周辺の風速場および気温場の数値解析

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1. 背景と目的
近年、ヒートアイランド対策が政策的課題に取り上げられる機会が増えていることから、各種対策の気温低減に対する有効度を定量的に明らかにすることが求められている。この種の解析を行う上で、都市キャノピーモデルは、必要となる計算機資源や取り扱い可能な都市構造の観点から、実用上最も有効な手法の一つであると考えられている。しかしながら、都市キャノピーモデルはこれまでに十分な検証が行われてきていらないのが現状である。特に温熱場における妥当性については殆ど検証されていない。さらに、都市キャノピーモデルを検証するためのデータベースも（特に温帯場については）殆ど存在していないものと考えられる。
そこで本研究では、都市キャノピーモデルの検証を行う準備として、規則的に配列された建物群周辺の風速場および気温場の数値解析を行い、水平方向に空間平均された気温及び風速のデータベースを構築した。さらに、数値解析結果を風洞実験結果と比較して精度検証を行った。

2. 計算概要
図1に示す計算領域において、k-εモデルによるCFD解析を行った。ここで、建物群は規則配列された70個の木製ブロックで模擬されている。尚、床面からの熱放射や伝熱の影響を考慮するために、ブロック壁面の温度を変化させた5ケースの解析を行った。

3. 結果
風速u（主流方向）と気温θ_aの鉛直方向の空間平均分布を図2に示す。鉛直高さ、横軸Z/Hは無次元化した鉛直高さ、横軸Hは無次元風速（図2左）および気温（図2右）である。尚、通路と建物を含む1ブロックの領域内において、空間平均分布を算出した（図2上）。
図2から、風速分布は、各ケース共に風洞試験データ（WTT data）と良く一致しているもの、気温分布は各ケースで差があることが分かる。本解析では、Case2、3、4が、風洞試験結果に比較的近い気温分布を再現している。
流れ場の詳細検討のため、通路側・建物背後側（図2上参照）の局所風速と局所気温の鉛直分布を図3、4に示す。局所風速分布は、通路側（図3a）・建物背後側（図3b）共に、風洞試験と良く一致し、建物背後側の逆流域（u/w∞<0）も再現されている。一方、建物背後側のキャノピーモデル内における気温分布（図4bのZH<1）は、全ケースで風洞試験より低下し、床面からの熱対流・拡散が十分模擬されていない。これは、建物背後側が、流速の極めて低い領域であり、今回使用した標準k-εモデルの適用範囲外になるためと考えられる。
今後、CFD解析の気温精度向上の検討と共に、これらの空間平均データを用いて、都市キャノピーモデルの検証・開発を行う予定である。
図 1 計算領域。

図 2 水平方向に空間平均された風速鉛直分布（左）と気温鉛直分布（右）。

図 3 局所の風速鉛直分布。

図 4 局所の気温鉛直分布。
メソスケール対流雲中での雲粒の衝突成長に及ぼす乱流の効果

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1. 緒言

気相乱流中で波滴粒子が互いに衝突するという現象は、工業装置内流れだけでなく環境中の流れの中でも広く見られる。最近の研究によれば、雲粒の衝突頻度が乱流によって増大させられることが明らかになっている(1)(2)。例えば、Lynn et al. (2) は Pen State-NCAR Mesoscale Model (MMS) を用いてフロリダで発生したスコールラインを計算し、雲粒の衝突成長に及ぼす乱流効果を調べた。その結果、乱流効果を考慮すると降雨の発生時期や発達をよりよく予測できることを明らかにした。彼らの研究は、実際の地形上で発達するメソスケール雲における乱流効果を調べ初めての研究である。しかしながら、彼らは雲粒衝突に及ぼす乱流効果を表現するために、乱流強度等の時間変化を考慮せずに単に衝突頻度を定数倍するという手法を用いた。このため、雲粒の衝突成長に及ぼす乱流効果が正しく捕らえられていない可能性があり、乱流効果が十分に解明された研究とは言えない。

そこで、本研究はメソスケール対流雲における乱流の効果を詳細に調べることを目的とした。そのために、まず乱流中での雲粒の衝突頻度を予測可能な衝突モデルを導入した。さらに、その衝突モデルを組み込んだメソスケールモデルを開発し、実際にメソスケール対流雲に適用した。

2. 雲粒の衝突成長に及ぼす乱流の効果

雲粒成長を考慮する手法として、雲粒を大きさごとにいくつか分けけて記録すると、この手法が開発される。その手法によって雲粒成長を計算する場合、衝突による雲粒の密度関数

\[
\frac{dn_f(r, x, t)}{dt} = \frac{1}{2} \int K_c(r', r'' n_f(r') r''(r''')dr'' - \int K_c(r, r') \int n_f(r)n_f(r')dr' 
\]

ここで、 \( r'' = (r^2 - r^1)^{1/3} \) である。また、 \( K_c(r_1, r_2) \) は衝突頻度因子と呼ばれ、半径 \( r_1 \) を持った粒子と半径 \( r_2 \) を持った粒子の衝突割合を表す。代表的な衝突頻度因子モデルとして、Hydrodynamic Kernel モデルがある。しかし、このモデルは気相乱流の効果を考慮できない。そこで、我々のグループは乱流の影響を正確に考慮できる衝突頻度因子モデルを開発した(1)。図 1 に開発した衝突頻度因子モデルの結果と 3 次元 DNS (直接数値計算、direct numerical simulation) によって得られた結果を比較して示す。図 1 から、本モデルと DNS の結果が良く一致することが分かる。

図 1 衝突頻度因子モデルの結果。
3. メソスケール対流雲に対する数値計算

本研究では、メソスケールの山岳降雨現象に対してシミュレーションを実行した。力学過程の計算には雲解像計算手法 CReSS の力学モデルを用い、雲物理過程の計算には前節で概説した独自開発モデルを用いた。

40 km×10 km×15 km の計算領域中央部に、高さ 1 km、半倍幅 2 km のベル型の山を配置し、湿潤飽和大気を一様流速 15 m/s で流入させた。水平方向には格子幅隔 200 m の等間隔格子を用い、鉛直方向には平均間隔 200 m であり地表付近で密となる不等間隔格子を用いた。計算領域側面には周期境界条件、出口境界に放射境界条件をそれぞれ用いた。図 2 に t = 5600 s における地表面でのスパン方向平均降雨強度を示す。なお、t = 5600 s には無次元時間 t' = Ut/la が 42 の時刻に相当し、その時刻では十分に設定の数値に達していると考えられる。図 2 より、RUN-T の場合、すなわち衝突に対する乱流効果を考慮した場合は、衝突に対する乱流の効果を考慮しない RUN-NoT の場合に比べてより風下の位置から降雨が始まることがわかる。これは乱流によって衝突成長が促進され、降雨開始が早まるためだと考えられる。また、主流方向に平均した降雨量は RUN-T の場合には 1.25 mm/h であり、RUN-NoT の場合の 1.05 mm/h に比せて約 20% 大きい。これは、乱流による衝突成長促進効果によって、より多くの降雨がもたらされることを意味する。

![図 2 スパン方向平均降雨強度。](image)

4. 結言

乱流中での粒子の衝突度を予測可能な新たな衝突度因子モデルを開発し、さらにそれを組み込んだ高解像度メソスケール気象モデルを開発した。開発した気象モデルを用いて山岳降雨現象に及ぼす大気乱流の影響を調べた結果、大気乱流によって雲粒の衝突成長が促進され、降雨量が増大されることが示された。

参考文献
二酸化炭素濃度の高解像度でのシミュレーションと
カラム濃度のシノプティック規模での変動

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1. 目的
本研究では緯度経度両方向に 0.25°以上の高分解能で全球的に走らせる事のできる大気微量成分輸送モデルを、これまでの NIES オフライン輸送モデルに改良を加えて開発した。高解像度モデルは、地表面フラックスの不均一性をより詳細に説明でき、大気汚染の進んだ地域（大気）と比較的きれいな地域（大気）の区別をより鮮明に行えるので、二酸化炭素輸送のシミュレーションを行う上で有益なものとなる。さらに、高解像度モデルは、これらの特徴は、全球あるいは局地的な二酸化炭素フラックスのインバースモデリング解析を行うにあたって、短期的（数時間から数週間）な変動をノイズではなくシグナルとして扱うための基礎を与えている。
また、シノプティック規模の気象変化が二酸化炭素の気柱平均濃度に与える影響も解析した。
鉛直対流モデルの誤差による影響を受けにくい為、カラム（気柱）濃度は大気中二酸化炭素のインバースモデル計算へのインプットとしても有用である可能性が高い。衛星観測は、地上観測よりもカラム濃度測定において有効であるかかもしれないが、より高い観測精度（1%以下）が求められる。晴天の場合においては、約 0.3～2.5 ppmv の精度が得られると考えられるが、雲のある天気の場合、衛星観測は困難になり、データの不確定性は増大する。

2. モデルと解析手法
NIES オフライン輸送モデルの時空間分解能に関するアルゴリズムにのみ改良を加え、モデルの水準方向の解像度を 2°、1°、0.5°、0.25°に設定し、二酸化炭素の地球規模での輸送をシミュレートした。このモデルでは、水平方向解像度が 1°×1°で鉛直方向が 26 レベルの NCEP の客観解析の風向き風速データをそれよりも高解像度の化石燃料燃焼による CO2 排出データと共に用いた。
ベクトル計算機能つきの NEC SX のような大規模な並列計算システムに対応できるようなプログラムを作成するために、まず地球データを緯度・経度に分割し、更に風向風速の客観解析データの前処理とフラックスの補間、そしてアウトプットの処理がそれぞれの緯度・経度内で（お互いの数値に影響を及ぼすことなく）行われるようにした。ノード間の情報交換を極力避け、各々の CPU が独立して計算することにより効果的な並列計算がなされるようにした。NEC SX を用いた実験計算ではプロセッサーごとの計算量、単独プロセッサーを使ったときに比べ、1～0.125°の水平分解能においては 60 ドメインで 70%に抑えられる事がわたった。地表面二酸化炭素フラックスとして、季節変動を含んだ海洋フラックス、日変動を含んだ Sib2 の陸域生態系フラックス、そして Marland et al. (2003) を基に高分解能化された化石燃料燃焼源のフラックスを用いた。境界層の混合過程をより明らかにするために、鉛直方向の層の数を 47 に増やし、また 3 時間毎に更新される PBL 高度のデータ（ECMWF）も併せて用いた。この鉛直方向の層の高度は高いほど厚く設定され、地表近くでは 12 hPa、対流層では約 25 hPa である。
シノプティック規模の気象変化が二酸化炭素の気柱平均濃度に与える影響を、気柱平均の二酸化炭素の偏差と地上気圧の偏差の相関関係を使い解析した。地上気圧を二酸化炭素カラムのダイナミックスと動的バイアスの指標とした。NIES 大気微量成分輸送モデル (2.5°×2.5°の解像度) と、様々な種類のフラックス（化石燃料燃焼源、陸域生態系（CASA モデルによる NEP）、海洋）そしてインバースモデルによる領域フラックス（陸域 11 と海洋 11 地域に全球を分割）を使って長期的な (1988～2003) 二酸化炭素のシミュレーションを行った。
3. 成果

図 1 に示すように、シミュレートされたシノプティックスケールの変動サイクルにはモデルの解像度による違いはあまり見られなかったが、高解像度モデルのほうが観測値により近い濃度を再現している。日変動する地表面フラックスが存在する場合、人為起源の都市からのプルームは、ある程度以上の解像度を持ったモデルによって、日変動の振幅が少ない地表レベル以上の高度においてのみ再現が可能であった。大都市から 300 km 以内に位置する地域の多くでは、解像度を 2.5°から 0.5°に向上することにより、大気汚染の進んだ地域とそうでない地域のコントラストがより鮮明にシミュレートされ、平均濃度に多大な影響をもたらした。

図 1 Pt. Barrow での地表面レベルの CO₂ の時系列変動。観測値と様々な分解能モデルのデータ。

シノプティック規模の気象変化の影響の解析では、夏季の低気圧域において、低い二酸化炭素カラム濃度がシミュレートされた。二酸化炭素カラム濃度の差と気圧の差異には夏の間の相関関係が見られ、西伯利亚と北米の一部においては高いところで 0.7 の正の相関関数が推定される。これにより、低気圧による上昇流により、夏の間の二酸化炭素濃度が低い PBL の空気が上昇し、その領域の二酸化炭素リサイクルが減少するからだと考えられる。逆に、冬のように、二酸化炭素の発生域が PBL の中にある場合、カラム濃度は低気圧によって高くなる。平均相関関数値は 1 年のうちに 0.6 から 0.6 の間で変動する。気圧との相関関係によって、二酸化炭素濃度のシノプティック規模での変動の大部分は説明がつくと考えられる。
積雲対流と大規模運動の相互作用についての数値実験：
雲活動の自発的集中化のメカニズム

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1. はじめに
地球の熱帯雲活動は、個々の積雲から始まって、メソシステム・クラウドクラスター・総観規模擾乱・感星規模の擾乱、そして気候学的構造に至るまで、種々の時空間スケールの構造を持つ。これらの構造のうち停滞性で大規模なもの、通常、海星温の非一様などの外的強制に支配されると考えられている。しかしこれらまでの研究により、その様々な外的条件が無くとも、一万キロを超える規模を持つ停滞性の雲活動集中化が起こる場合があることがわたった。今年は、この雲活動集中化のメカニズムを調べた結果を報告する。使用したモデルは非弾性方程式系に基づく水平鉛直の2次元雲対流モデルであり、計算領域は鉛直約23 km、水平には32,768 km（周期境界条件）である。計算は約30日行った。

2. 結果
これまでの研究により、大規模な停滞性雲活動の存否は放射冷却の鉛直構造に敏感であり、冷却が上層で強い場合に顕著になる。この選択性はwave-CISKによって伝播性雲活動の存否(Hayashi, 1970)と同じである。Nakajima(2002)は、放射冷却が中層で弱き下層と上層で強い場合にも伝播性雲活動が起こる事を示した。そこで、この場合についても停滞性の大規模集中化が起こるかを調べた。その場合、図 1 に示すように、波長数万キロの伝播性組織化に重なって停滞性的波長3万キロの雲活動の濃淡が成長してくる。波長3万キロの伝播性組織化と波長3万キロの停滞性構造がなぜ共存するかは依然として問題である。wave-CISKの線形論（波長は予想できない）を再検討してみると、下層の鉛直流と雲加熱の結合が“中程度”的場合に伝播性不安定擾乱が、非常に強い場合には停滞性不安定擾乱が生じることがわかる。そこで、数値計算の結果を水平方向のスペクトル分解し、降水量（雲の加熱に対応する）と下層鉛直流の対応の波長依存性を調べた。その結果、図 2 のように、降水量の振幅と鉛直流の振幅はほぼ比例し、その比例係数は長い波長ほど大きいことがわかった。これは線形論の予想と対応する。

図1 モデル内の降水量分布の時間発展。横軸は、モデル内の水平方向（32,768キロ）、縦軸は計算時間（下向きに進む）を示す。
図2 数値モデルにおける降水量と鉛直流（高度1.5キロ）の結合定数の水平スケール依存性（横軸：水平波数、縦軸：比例係数）。緑のマーカーが上層で放射冷却が強い場合（停滞性大規模組織化が生じる）、赤のマーカーは下層で放射冷却が強い場合（停滞性大規模組織化は起こらない）。
二重拡散流体中の熱と塩分の差分拡散

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1. 背景と目的
海洋の流れは、海水密度を決定する熱と塩分の分布に大きく依存する。海洋数値モデルにおいては、通常、熱と塩分の乱流拡散係数は等しく置かれて来た。これは、レイノルズ数が十分大きくなり、分子拡散の効果は非線形効果に覆い隠されるという仮定のもとでの話である。しかし、近年、両者の違いの可能性やその影響が議論されるようになった。海洋観測や室内実験においても両者の違いを指摘する結果が現れ、IAPSOでも拡散係数に関するワーキンググループを設置した。室実実験によれば、差分拡散は、海洋観測でよく用いられる浮力レイノルズ数 $Re = N^2 / f$ が10以下程度で発生する（$e$ は乱流のエネルギー散逸率、$v$ は流体の動粘性率、$N$ は浮力振動数）が、この値は実際の赤道や極付近の密度変動層、海洋深層において典型的な値である。本研究では、熱と塩分の分子拡散の違い、すなわち、diffrential diffusion（以下、和訳として「差分拡散」と呼ぶことにする）が、それらの乱流（渦）拡散係数に与える影響を数値シミュレーションにより解析する。なお、実際の系では熱（温度）のプラントル数は $Pr_t = 6$ 、塩分のプラントル数は $Pr_s = 600$ であるが、計算機の能力の限界のため、ここではそれぞれ、$Pr_t = 1$ 及び $Pr_s = 6$ として計算した。

2. 結果及び結論
差分拡散の存在自体は、流れの渦構造のような小スケールの構造にあまり影響を与えない。図1は、渦の等価面であるが、こうしたpan-cake（パンケーキ）構造自体は、密度成層を構成する要素が単に同（温度）だけの場合と実は大きな差はない。実用上重要な問題は、大スケールの構造が強く熱と塩分のフラックス（あるいは乱流拡散係数）である。これらは、比較的初期の時刻に分子拡散の影響による差を生じる。しかしその差は、時間が経っても消減せず、ずっと後まで影響を引きずることになる。図2は、塩分と熱の渦拡散係数（$K_s$ と $K_t$）の比 $d = K_s / K_t$ の時間変化を示しているが、時間と共に減少し、0.5付近の小さい値に漸近することがわかる。このことは、$d = K_s / K_t = 1$ を仮定すると、拡散係数が實際とずれてくる可能性を示している。ただし、渦拡散係数も密度成層の強さに依存するため、浮力レイノルズ数に依存し、室内実験では $d = 0.2$ 程度の低い値も得られている。したがって今後、観測等との比較の観点からも、浮力レイノルズ数依存性の解析が不可欠である。

図1 成層流体中の渦度の等価面（パンケーキ構造）。
図2 塩分と熱の渦拡散係数の比 $d$ の時間変化。
水惑星実験における赤道域降水量の多様性:
鉛直乱流混合過程の影響

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1. 背景と目的

本研究の背景となる我々の目的は、数値実験を通して大気大循環に内在する力学的構造を抽出
して理解し、大気大循環モデルに表現されるべきさまざまな過程とその表現方法に関する知見を
得ることにある。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に適したモデ
ル群を通し連なるソフトウェア群を作成構築することである。第二は、力学的構造の抽出を試みる
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さを削除することにより、大気大循環に関与する基本的な力学的特性が抽出
できるのではないかと期待するわけである。我々はそのような単純な設定での実験として水惑星
実験（地表面から陸地をとりきり海洋だけにした地球上の循環を数値的に調べる実験）を行って
きている。特に赤道域での降水量と循環構造の形成に注目し、数値モデルで得られる降水量
パターンの多様性を探索し、その発現に努めた。ここでは赤道域間接降水構造の形成に対する
鉛直乱流混合過程がもたらす影響を、鉛直乱流混合過程を単純化することにより調べる。

2. 赤道域降水パターンの成因と鉛直乱流混合

上層冷却型の放射冷却分布を用い、積雲パラメタリゼーションとしてKuoスキームを用いた
場合に対し、鉛直乱流混合を単純化した実験を行った。Kuoスキームはwave-CISKの力学が働
きやすいと期待される積雲パラメタリゼーションであり、実際の降雨実験（K-UC実験）によれば、対流層上層で強い冷却を与えると格子点スケールの降水域の
東進が観測される。しかし、その東進パターンはHayashi and Sumi（1986年）や
Nagamuti and Hayashi（1991年）とは異なり、赤道円周上に強い降水域が一箇所しか存在しないようなパターンとな
っており（図1）。この原因を探るべくモデルの物理過程を順次単純化し、線形wave-CISKモデ
ルに近づける操作を行った。その結果、乱流鉛直輸送過程を単純化し、地表面フロックスを与える
るパルク係数を安定度やシアーや依存しない定数とし、鉛直乱流混合も安定度やシアーや依存し
ない固定鉛直拡散係数で与えることにより、強い降水域が赤道円周上に複数発生する傾向
が現れた（図2）。鉛直輸送過程は海面から境界層を経て自由大気への水蒸気供給を支配している。
現在のモデルではHayashi and Sumi（1986年）やNagamuti and Hayashi（1991年）に比べて循環場が鉛直
輸送過程で強くフィードバックし、降水が起こりにくくなっており、一度降水が発生すると強い
降水になるためますます他の降水の発生に対して抑制的になるのだと想像される。

3. GCMの階層的モデル化: DCPM

線形wave-CISKモデルから大気大循環モデルに至るまで、構成素過程の複雑度を順次上げて
いったモデルを階層的に用意し実験を行うことができれば、大気大循環モデルにおける降水パタ
ーンの形成をよりよく理解しコントロールすることができるだろう。現在、物理過程の着脱を簡
便に実行するために、新しく大循環モデルを組み直し（惑星大気モデルDCPAM）、またそのため
の関連するソフトウェア環境の構築を進めている。今年度は、物理過程の実装（AGCM5.3と同様
の乾燥対流調節、湿润対流調節、大規模凝結、放射（4色簡易バンドモデル）、鉛直拡散（Yamada

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and Mellor level 2.5)、バルク法を用いた地表面フラックス）を進め、水経系実験の追試を試験的に開始した。支援ツールとして、Fortran90モジュール・サブルーチン・関数の機能や引数を記述したリファレンスマニュアルを自動作成するRubyユーティリティRDocを改良作成した。

図1 K-UC実験における赤道断面図のスナップショット。 (上) 凝縮加熱 (K s⁻¹)、 (中) 湿度 (kg kg⁻¹)、 (下) 温度 (K)。対流圏の湿度異常が波状1的であることに注意。

図2 地表面フラックスを与えるバルク係数と鉛直乱流混合を定数にした実験における赤道断面図のスナップショット。 (上) 凝縮加熱 (K s⁻¹)、 (中) 湿度 (kg kg⁻¹)、 (下) 温度 (K)。対流圏の強い高温異常が複数存在することに注意。
進行波上の気流中の運動量輸送とスカラー輸送の相似性

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1. はじめに
海洋から大気へ輸送される波熱は、熱帯低気圧のエネルギー源となる。したがって波熱や顕熱の交換過程を理解し、これらの交換量を正確に見積もることは、気象・海象を正確に予測する上で重要である。これまで、鉄塔や船による現地観測を実施することにより、大気と海洋間の波熱・顕熱の交換量が計測されてきたものの、波熱・顕熱の交換に対して水面波が与える影響を詳細に調べた研究は他には少なく、運動量交換に関する理解に比べて、その知見は乏しい。本研究では、3 次元直接数値計算（DNS）を実施することにより、水面波上の気流中における運動量輸送及びスカラー輸送の波縁依存性を調べ、運動量輸送とスカラー輸送の関係について議論する。波縁（c/uₙ）を 0.1 までの 10 ケースについて計算し、スカラー輸送・運動量輸送の波縁依存性を調べることにより、その相似性を議論する。

2. 数値計算の概要
本研究で考える流れは、下面に 2 次元的な進行波面、上面に水平な摩擦なしの境界をもつ領域内の気流である。水平方向は周期境界条件を仮定する。進行波と同じ速度で移動する座標系を用いて計算するため、進行波の形状は時間的に変化しない。また流れは、一定の水平方向圧力勾配によって駆動されるとし、完全に発達した定常乱流に対する計算を行う。基礎方程式は、非圧縮性流体に対するナヴィエ・ストークス方程式、連続の式及びパッシブスカラーの輸送式である。波形勾配を ak = 0.1、レイノルズ数 Reₜ(= uₙh/ν) を 150、波縁 c/uₙ は 0.2, 4, 5, 6, 8, 10, 12, 16 及び 20 の 10 ケースについて計算し、乱流統計量が定常になるまで時間発展させた。

3. 結果
対数域での平均流速及びスカラーの波縁との関係を見るために、運動量輸送及びスカラー輸送に対する無次元粗度高さ z₀₉ₚ（= z₀ₙg / uₙ²）及び z₀ₙ₉ₚ（= z₀ₙg / uₙ²）を図 1 に示す。図 1 は、z₀ₚ₉ₚ と z₀ₙ₉ₚ に同様の波縁依存性が見られ、その波縁依存性が同じ程度の強さであることを示している。このことは、進行波上の気流中におけるスカラー輸送と運動量輸送との間に強い相似性があることを意味している。

![図 1 運動量とスカラーの無次元粗度高さ。○：スカラー、●：運動量。]
準地衡風渦領域の統計性

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1. 緒言
大規模な地球流体運動は、地球の自転と密度成層効果のために、近似的には鉛直直上での異なる層ごとに二次元運動とみなせ、準地衡風近似で記述される。本研究では、連続渦度分布を離散点渦系で近似してその「統計力学」を調べる。準地衡風近似を用いて散逸過程を伴わない点渦系と散逸過程（渦の合体）をモデル化した渦渦系のそれぞれの統計的性質を理論・数値計算の両面から調べた。

2. 準地衡風点渦の統計力学
無限の領域中のある領域に多体の点渦をランダムに一様分布させ、点渦系の統計的性質を調べる。数値計算は、渦数 \( N = 2000 \)、循環 \( \Gamma_{1,2,\ldots,N} = 0.5 \)、エネルギーは \( E = H/N^2 \) = 0.123 として行った。エネルギーはランダム初期渦分布を \( 10^6 \) 個用意したとき系が取り得る総エネルギーの頻度が最も高かったものを用いた。この点渦系は無次元時間 \( t = 10 \sim 20 \) で平衡状態（図1）となるため、\( t = 20 \sim 200 \) を時間平均して平衡分布を求めた。平衡状態における渦の分布は渦重心に対して軸対称であるため、周方向に平均を取り、渦重心からの距離を \( r \)、鉛直方向の座標 \( z \) とした平衡分布 \( F(r, z) \) を調べた。図2のように、\( |z| \leq 1.459 \) (全体の約 60%) ではほぼ同じ平衡分布であるが、\( |z| > 1.459 \) では \( |z| \) が大きくなるほど平衡分布は渦重心に集中した。中心部 \( |z| \leq 1.459 \) と上下部 \( 1.459 \leq |z| \leq 2.432 \) に現れた平衡分布の差を「End-Effect」とおよび、それを検証するために「一層」 \( \langle z_{1,2,\ldots,N} = 0 \rangle \)、「二層」 \( \langle z_{1,2,\ldots,N} = -h/2, z_{N+1,N+2,\ldots,N} = h/2 \rangle \) に配置した場合について数値計算を行った。「一層」の場合、平衡分布 \( F(r) \) の最大エントロピー理論に基づく理論解を求めたところ数値計算結果とよく一致した。エネルギーを変化させると、エネルギーの増加とともに平衡分布は渦重心 \( r = 0 \) に集中することがわかった。「二層」の場合も平衡分布 \( F(r) \) の理論解を求めたところ数値計算結果とよく一致した。

![図1 分布。(t = 0,200)](image1)

![図2 平衡分布。F(r, z)](image2)
3. 球渦モデルの統計性

散逸過程をモデル化するために、点渦に半径を持たせた球渦モデルを導入する。直接数値計算である CASL 法では、水平方向の中心間距離が 2 つの球渦半径の 1.3 倍、鉛直方向には球渦半径の和より近付くと合体し、エネルギーとエントロフィーが減少した合体後の渦が合体後に形成される。エントロフィー散逸係数 \( F_s = 0.988 \) で整合性のある結果が得られた。エネルギー・エントロフィー・渦数はある時間領域においてべき乗則で減少する。渦度分布は時間発展とともに一様となり、一様渦度領域 (Vortex-Patch) の形成が確認された。

4. 結言

本研究では準地衡風点渦、準地衡風球渦の統計性を調べ、次の結果を得た。

- 準地衡風点渦
  「一層」、「二層」ともに最大エントロピー理論を用いて求めた平衡分布は数値計算結果とよく一致した。また、最大エントロピー理論による解析でエネルギーを上、下に変化させると、エネルギーの増加とともに平衡分布は渦重重に集中した。
  「連続層」の平衡分布は、鉛直方向において中心部が 60% の領域では変化が見られなかったが、上(下)部では端ほど渦重重に集中した。
- 準地衡風球渦
  合体にともなってエントロフィーを散逸させる球渦モデルでは、散逸係数 \( F_s = 0.988 \) の場合、エネルギー・エントロフィー・渦数の減少率がべき乗則に従った。
  渦度は時間発展とともに一様となり、一様渦度領域 (Vortex-Patch) の形成が確認された。
比較的高解像度の大気大循環モデルに見られるアンサンブル揚動の特徴について

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1. はじめに
数値天気予報が外れる主な原因として、一般に、初期値に含まれる誤差、予報モデルの不完全性、大気の不安定性の三つが挙げられる。最初の二つは観測の充実化や数値モデルの開発・改良により改善されてゆくものと考えられるが、三番目は初期値の鋭敏な依存性や「カオス」という言葉で表される大気固有の性質であり、直接的に対処する手段は存在していない。大気の不安定性の特徴をより深く理解することが、数値天気予報だけでなく、大気モデルを用いた今後のシミュレーション研究において重要と考えられる。本研究では、比較的高解像度の大気大循環モデルによるアンサンブル数値実験の結果を用いて、模型の中で発達する微小な揚動の空間構造の特徴について調べた。

2. 成長モード法の揚動
空間解像度 T59L48（水平解像度 80 km、鉛直 48 層）の大気大循環モデル AFES（AGCM for the Earth Simulator; Ohtuchi et al., 2004）を用いて行なった成長モード法によるアンサンブル数値実験の揚動について解析を行なった。成長モード法のコントロールランを 2004 年 3 月 1 日から 4 月 10 日までの現実大気の時間発展に相当するようナッジ法により強制し、その後のコントロールランに沿って 99 個の揚動を 12 時間毎に直交化と規格化を繰り返しながら成長させた。規格化による揚動の大きさは、風速でみて 1 m/s 程度としている。この実験から現れる 99 個のリアプノフ指数は全て正値であり、モデル内で成長する揚動の全体を把握するためには、より多くのメンバーで構成されるアンサンブル実験が必要であることを示唆している。この実験の成長モードの揚動は、中・高緯度帯で大きな振幅を持ち、その空間スケールはオーダー 1000 km の総観規模スケールよりもやや小さい。これらの揚動は中緯度帯の低気圧の発達に伴って生成された局所的に不安定な領域で急速に成長したものと思われる。

3. 考察
本研究の結果は、一つの試験的な成長モード法の実験の解析結果に過ぎない。規格化をする時間間隔や規格化の大きさなどの成長モード法の実験設定を変えることにより、得られる揚動の特徴が変化することが考えられる。例えば、規格化の大きさを小さくすることにより、本実験の中では非線型効果の寄与により飽和状態に達していたと思われる地形性の不安定や対流性の不安定が、成長モードの揚動の中に支配的な構造として出現する可能性が考えられる。本研究で得られた結果を基に、実験設定を変えた数値実験を系統立てて行ない、その中で発展する揚動の特徴を調べることにより、大気不安定性の理解を深めてゆく計画である。
流域土砂動態モデルの適用による久著呂川流域の土砂収支解析

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1. はじめに
近年、土砂災害防止、河川環境整備、海岸保全のために、流域源頭から海岸までを一体として定義された『流域』の総合的な管理の必要性が唱えられ、これに対応した形で、上流域山地河川における土砂動態を予測する土砂水理学に基づく機構モデル、流域全体に対するモデルが提案されている。日本の場合、上流域から中流域への移行に伴い、中間山地帯の農林業が出現する。こうした地域での土砂生産は上流域地域での土砂生産と相まって、より下流域での土砂問題を引き起こす可能性がある。本研究では対象とする釧路湿原に流入する久著呂川流域（図1）では1960年代後半から80年代前半にかけて、流域内農業開発が進められた。上流域では伐採が容易な領域での伐採の結果、パッチ状の林業帯が出現した。中流域では森林伐採後、農地化が進められた農地の保全と洪水災害防止のため、部分的な河川改修が行われた。下流域では湿原を埋め立てるとともに農業排水路網整備、蛇行河道の固定化と直線化が進められた。その結果、林業の減少と農地拡大との相乘効果で増大した流出土砂は、通行能力の向上した河川を通じて釧路湿原に流入、氾濫堆積し、湿原の乾燥化という問題を引き起こしている（図2参照）。本研究では、湿原管理にとって必須となる久著呂川流域での土砂動態予測と『流域』における河川流域内農林業地域での土砂動態予測のため、土砂流出モデルの構築を行い、実現したモデルに基づき、久著呂川流域での土地利用の流下方向変化にともなう土砂生産量変化についての検討を行った。

図1 久著呂川流域と観測点、(St.1～St.4) 図2 久著呂川断面勾配。

2. モデル概要
著者らは、水による物質輸送の概念に忠実であること、地形特性、土壌特性、土地利用状況といった流流出特性に及ぼす様々な要因の空間分布の取り込み易さの観点から、河川水系網において合流・分流を端点とする河道区間への集水域を単位流域としてモデル化し、これらを多数接続させることで流域全体を表現するベクトル型のモデルを用いている。著者らは、対象流域の環境情報のデータベースを前提として、流域面での(i)降雨流出モデル、(ii)土砂生産場から河道への土砂輸送モデル、(iii)河川流モデル、(iv)河道での流砂モデルの4つのサブモデルより構成される流域土砂動態モデルを提案している。
3. 結果
図3は本文モデルであるStanford Modelと河道網モデルとしての拡散波モデルとの結合モデルによる下水道流量観測点での河川流量に関する再現計算結果であり、十分に傾向を表現している。
図4はSt.1でのwash loadの濃度の時間変化であり、計算結果は急激に立ち上がり、急激に減少する傾向をとらえている。これは特に河岸侵食がarmor coatの破壊時、短時間に起こるためで、他の成分は低濃度で継続するものと考えられる。
図5は、9月27日～9月30日の降雨・出水による久著呂川での伐採地・農地・河岸の3領域からの土砂生産量と各河道区分での平均河床高変化を示している。計算された洗掘・堆積の経年方向の傾向は、従来の報告と類似している。また、計算結果によると、農地由来の微細土砂量は全微細土砂量の約40%を占め、これは従来の観測値をもとに推定した20～30%の2倍程度である。土地利用との関連では、上流域では伐採地と河岸が生産源となり、農地面積の増加とともに、農地が主な生産源になるものの、保全対策の増加とともに減少する。また、河道の直川路化・落差工設置の改修区間では、中流域ではあるものの河岸が土砂の生産源となっている。これらの結果は種々の仮定に基づいており、得られた結果は、一種の土砂生産ポテンシャルを表現しているものと判断される。
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及び応用(2)

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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 (～100 m) 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、適応型数値的手法を提案し、安定、効率的および正確な高解像度の大気地域数値モデルの開発及び応用を行う。

2. 概要
これまで我々は斬新な大気局地数値モデルの開発・応用を行っている。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、非常常三次元可圧縮流体を計算するための SIMPLER (Semi-Implicit Method for Pressure-Linked Equation Revised) アルゴリズムを採用した。我々は、慣例となっている地形追隨規格化(terrain-following normalization)は行わず、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、ブロッキング・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、完全時間陰解法(fully time-implicit scheme)を利用している。ここで都市乱流計算の事例として、都市キャノピー層内の流れシミュレーションを本モデルで実行した。シミュレーションの結果により都市キャノピー層の気候特性がよく再現されたことがわかった。

3. 成果
図1は都市キャノピー及び計算領域を示している。図2は大気不安定層の場合に都市キャノピー層内の水平面温度と速度ベクトルが描かれている。図からは大気不安定層時、都市キャノピー層の熱・流れの様子が詳細に示されている。

![図1 都市キャノピー及び計算領域。](image1)

![図2 都市キャノピー層の温度・速度ベクトル。](image2)
Program of
the 14th Supercomputer Workshop

October 30, 2006
National Institute for Environmental Studies
Tsukuba, Ibaraki, Japan
Program of the 14th Supercomputer Workshop

11:00〜17:20, October 30, 2006
National Institute for Environmental Studies, Tsukuba, Japan

11:00〜11:05 Opening address
Yasuhiro Sasano (Center for Global Environmental Research, NIES)

11:05〜11:15 Briefing
Yasumi Fujinuma (Center for Global Environmental Research, NIES)

11:15〜11:30 Simulation of the atmosphere in glacial age by general circulation model-Mechanism of atmospheric circulation change over East Asia and North Pacific
Wataru Yanase (Center for Climate System Research, University of Tokyo), et al.

11:30〜11:45 A study of aerosol impacts on climate using CGM
Daisuke Goto (Center for Climate System Research, University of Tokyo), et al.

11:45〜12:00 Lunch

13:00〜13:15 Numerical simulations of quasi-geostrophic vortex based on CASL algorithm and point-vortex method
Yusuke Hori (University of Electro-Communication), et al.

13:15〜13:30 Differential turbulent diffusion in the salt-heat system
Hideshi Hanazaki (Kyoto University)

13:30〜13:45 Effect of underground urban structures on hydrologic and nutrient budgets in Tokyo metropolitan area
Tadanobu Nakayama (NIES), et al.

13:45〜14:00 The effects of rainfall and wind shear on air-sea CO2 transfer
Satoru Komori (Kyoto University)

14:00〜14:15 Equatorial precipitation patterns in an aqua-planet experiment: effects of vertical turbulent mixing processes
Yoshi-Yuki Hayashi (Hokkaido University), et al.

14:15〜14:30 Numerical simulation of temperature distribution within and above urban canopies
Yasunobu Ashie (Building Research Institute), et al.

14:30〜14:45 The flow of program turning of a supercomputer
Ryuji Tsukamoto (NEC)

14:45〜15:05 Coffee Break

15:05〜15:20 Indirect acidification in East Asia: enhanced deposition of semi-volatile aerosol components associated with changes in the gas-aerosol partitioning
Mizuo Kajino (Disaster Prevention Research Institute, Kyoto University), et al.
15:20～15:35  Numerical experiment on the large-scale organization of cumulus convection
            Kensuke Nakajima (Kyushu University), et al.

15:35～15:50  A comparison of numerical simulation and wind tunnel experiment on unstable
            stratified flow within and above a modeled urban canopy
            Weiming Sha (Tohoku University), et al.

15:50～16:05  Past 25-year simulation of stratospheric ozone with a chemistry-climate model
            Kiyotaka Shibata (Meteorological Research Institute), et al.

16:05～16:20  Spatial structure of perturbations growing in an atmospheric general circulation
            model
            Shouzo Yamane (Chiba Institute of Science)

16:20～16:40  Inferring CO₂ fluxes at regional scale by inverse modeling and using backward
            atmospheric tracer transport
            Claire Carouge (Center for Global Environmental Research, NIES), et al.

16:40～17:00  A future ozone layer prediction using CCSR/NIES Chemical climate model with T42
            horizontal resolution
            Hideharu Akiyoshi (NIES), et al.

17:00～17:15  Discussion

17:15～17:20  Closing address
            Yukihiro Nojiri (Center for Global Environmental Research, NIES)
スーパーコンピュータによる地球環境研究発表会（第14回）プログラム

日時：平成18年10月30日（月）11:00～17:20
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
笹野 泰弘（国立環境研究所 地球環境研究センター長）

11:05～11:15 スーパーコンピュータ利用研究概要紹介
藤沼 康宜（国立環境研究所 地球環境研究センター環境情報研究推進室長）

11:15～11:30 大循環モデルを用いた水期の大気場のシミュレーション—東アジア・北太平洋域の大気変動メカニズム
○柳瀬 壮1・阿部 彩子12（東京大学 気候システム研究センター、海洋研究開発機構 地球環境フロンティア研究センター）

11:30～11:45 GCM を用いたエアロゾルの気候への影響に関する研究
向井真木子1・五藤 大輔1・竹村 俊彦2・中島 映至1（東京大学 気候システム研究センター、九州大学 応用力学研究所）

11:45～13:00 Lunch

13:00～13:15 CASL 法と点滅法による気地衛風洞の数値シミュレーション
○堀 祐輔・星 伸太郎・李 英太・宮嶋 武（電気通信大学大学院 電気通信学研究科）

13:15～13:30 熱塩二重拡散系における乱流拡散係数
○花崎 秀史（京都大学大学院 理工学研究科）

13:30～13:45 東京都心部の地下構造物が水・物質循環に及ぼす影響について
○中山 忠昭1・丹治 三則2・渡辺 正孝12・盛岡 通3（国立環境研究所 アジア自然共生研究グループ、慶應義塾大学 環境情報学部、大阪大学大学院 工学研究科）

13:45～14:00 大気・海洋間の CO2 輸送に及ぼす降雨およびウインドシアの影響について
○小森 悟（京都大学大学院 理工学研究科）

14:00～14:15 水深実験における赤道域降水パターンの多様性：鉛直乱流混合過程の影響
○林 祥介1・石渡 正樹2・山田 由貴子1・森川 靖大1・高橋 芳幸1・中島 健介3・小高 正樹1・竹広 真一4（北海道大学大学院 理学院、北海道大学大学院 地球環境科学研究院、九州大学大学院 理学研究院、京都大学 数理解析研究所）

14:15～14:30 都市キャノピー内及び上空の熱的機構に関する数値解析
一ノ瀬 俊明1・足永 喜信2・河野 孝昭2・東海林 孝右1（国立環境研究所 地球環境研究センター、建築研究所 環境研究グループ）

14:30～14:45 コンピュータシステムの現状と将来
○塚本 龍治（日本電気（株）第一官庁システム事業部）

14:45～15:05 Coffee Break

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15:05～15:20 東アジアにおける半揮発性エアロゾルのガス＝エアロゾル分配変化を通じた間接的酸性化効果
○梶野 塚王1・植田 洋匡2（1京都大学 防災研究所、2(財)日本環境衛生センター 酸性雨研究センター）

15:20～15:35 積雪対流と大規模運動の相互作用についての数値実験：雲活動の自発的集中化のメカニズム
○中島 健介1・小巻 正嗣2・杉山 耕一郎2・北守 太一2（1九州大学大学院 理学研究院、2北海道大学大学院 理学部）

15:35～15:50 数値シミュレーション及び風洞実験による都市キャノピー層の不安定成層流の研究（A comparison of numerical simulation and wind tunnel experiment on unstable stratified flow within and above a modeled urban canopy）
○余 偉明1・阿部 敏雄2・足永 冨信2（1東北大学大学院 理学研究科、2建築研究所）

15:50～16:05 成層圈オゾン層の過去25年の化学－気候モデルによるシミュレーション
○柴田 清孝・出井 真（気象研究所 環境・応用気象研究部）

16:05～16:20 大気循環モデルの中で発散する擾乱の空間構造について
○山根 省三1,2（1千葉科学大学 危機管理学部、2海洋研究開発機構 地球環境フロンティア研究センター）

16:20～16:40 Inferring CO2 fluxes at regional scale by inverse modeling and using backward atmospheric tracer transport（インバースモデルと大気化物質逆輸送を用いた領域レベルでの二酸化炭素フラックスの見積り）
○Claire Carouge1, S. Maksyutov1,2, P. Peylin3, P. Bousquet1, P. Rayner3, P. Ciais3, T. Machida1, K. Shimoyama1, M. Arshinov4, O. Krasnov4, B. Belan4, G. Inoue4,5（1NIES, 2PRCGC/JAMSTEC, 3LSCE, Gif sur Yvette, France, 4,5Inst. Atmospheric Optics, Tomsk, Russia, 6Nagoya University）

16:40～17:00 CCSR/NIES T42化学気候モデルを用いたオゾン層の将来予測実験
○秋吉 英治1・吉野 宗佳1・永島 達也2・高橋 正明3・今村 隆史1・黒川 純一2・
汲川 雅之4・L.B. Zhou4・坂本 圭4（1国立環境研究所 大気圈環境研究領域、2国立環境研究所 アジア自然共生研究グループ、3東京大学 気候システム研究センター、4海洋研究開発機構 地球環境フロンティア研究センター）

17:00～17:15 総合討論

17:15～17:20 閉会挨拶
野尻 幸宏（国立環境研究所 地球環境研究センター副センター長）

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