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Foreword

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

In March 1992, CGER installed a supercomputer system (NEC SX-3, Model 14) to facilitate research on global change. This machine was subsequently replaced by a newer supercomputer (NEC Model SX-4/32) in March 1997. Proposed research programs are evaluated by the Supercomputer Steering Committee consisting of leading Japanese scientists in climate modeling, atmospheric chemistry, oceanic circulation, and computer science. After project approval, authorization for system usage is provided. In fiscal year 2002 (April 2002 to March 2003), 18 research programs were approved to use the supercomputer system.

The Supercomputer Activity Report Vol. 11 compiles the research results in fiscal year 2002. The research papers in this report do not necessarily show the final results of the research programs, which are to be published in the form of “full papers” upon completion of each research program. This report consists of research papers classified into four categories—Climate Modeling, Atmospheric and Oceanic Environment Modeling, Geophysical Fluid Dynamics, and Other Researches—together with the Overview of the Supercomputer Systems. We hope this report provides useful information on the global environmental research. In order to promote the exchange of ideas and opinions amongst the scientific fraternity about utilizing this supercomputer, the Research Integration Section of CGER would greatly appreciate any comments or suggestions on this publication.

In March 2002, we finished installing a newer supercomputer system (NEC Model SX-6) to replace the SX-4/32. Now the SX-6 is operating at its full capability due to the increase of CPU utilization, and we are considering ways and means to practice more effective operation. Although NIES was transformed from a research institute of the Ministry of Environment to an independent research agency on April 1st 2001, we will continue to support global environmental research.

January 2004

Shuji Nishiooka  
Executive Director  
Center for Global Environmental Research  
National Institute for Environmental Studies
Preface

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

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

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

January 2004

Gen Inoue
Director
Center for Global Environmental Research
National Institute for Environmental Studies
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1. Climate Modeling
A Study on Polar Ozone Destruction due to Bromine Species

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Abstract

Polar ozone destruction due to bromine species is studied using the CCSR/NIES nudging Chemical Transport Model (CTM). A new version of the CTM was developed, which includes BrOx and SOx chemistry as well as O³, H²O, NOx, CHx, and ClOx chemistry for stratosphere. An important property of bromine species is that bromine species are more easily photolysed than chlorine species. The ozone destruction effect through this photolysis property is studied in connection with the heterogeneous reaction and polar vortex condition. The years of 1995, 1996, 1997, and 2000 were investigated. It is shown that heterogeneous reaction processes produce a rapid increase in BrO in spring before ClO decreases, and that a strong downward motion in the Arctic polar vortex causes an increase in the total reactive bromine (Brγ) abundance, hence BrO abundance in the spring lower stratosphere. It is also shown that the bromine chemistry enhances the ClO dimmer ozone destruction catalytic cycle through the formation of BrCl.

Keywords: Nudging CTM, Ozone destruction, Bromine species, Photolysis, Polar vortex

1. Introduction

Some of the halon gases that release bromine atom in the upper troposphere and lower stratosphere have still been increasing in the atmosphere, although the amount of CFC gases and reactive chlorine in the lower stratosphere has already been at or near the peak (WMO, 2003). Yung et al. (1980) pointed out the importance of atmospheric bromine chemistry and the catalytic destruction of ozone by the ClO-BrO cycle. McElroy et al. (1986) showed that bromine was important for the formation of the ozone hole. Lary (1996) and Lary et al. (1996) studied the effect of bromine species on ozone depletion in the stratosphere and upper troposphere. They showed that in the polar region, formation of BrCl through the heterogeneous processes accelerates ozone depletion. Millard et al. (2003) estimated ozone destruction by the ozone destruction catalytic cycles in the northern high and mid latitudes in 1990s. They showed that the BrO-ClO catalytic cycle made a large contribution to the ozone destruction. Recently, Rex et al. (2003) reported that chemical models failed to reproduce the Arctic ozone destruction observed in January. They suggest that ozone destruction due to bromine species in the polar twilight could be one of the possible causes for the unknown part of the ozone destruction. Bromine species is photolysed more easily, thus sensitive to twilight of the solar radiation in the polar region. In this report we focus on the role of bromine to ozone destruction through the photolysis property. Heterogeneous reactions and a polar vortex condition are one of the factors that can make an effect on the ozone destruction through the photolysis.
2. Model Description

A nudging CTM has been developed at NIES using the CCSR/NIES Atmospheric General Circulation Model (AGCM). The AGCM was originally developed by Numaguti et al. (1997). The first version of the AGCM with coupled chemistry was described by Takigawa et al. (1999). The chemistry-radiation coupling scheme for a chemical-radiative coupled 1-D model developed by Akiyoshi (1997) was applied. The model has 30 vertical atmospheric layers. The top level is at an altitude of around 70 km. The improved points from the first version to the present version are as follows: (1) The gas phase chemistry scheme including bromine species developed by Akiyoshi (2000) were incorporated into the 3-Dimensional CCSR/NIES AGCM. The heterogeneous reaction scheme for a box model version of the SLIMCAT model (Sessler et al., 1996) was also incorporated. The scheme includes heterogeneous reactions on Polar Stratospheric Clouds (PSCs) of Supercooled Ternary Solution (STS), Nitric Acid Trihydrate (NAT), and ice. The sulfate chemistry to simulate H$_2$SO$_4$, sulfuric aerosols, and STS, was also included according to Takigawa et al. (2002). For simplicity, however, SO$_3$ is assumed to be in a photochemical equilibrium. Concentration of OCS and SO$_2$ were specified at the surface. These two are the source gases of the sulfur species in the model. The chemical schemes for the stratosphere are extended down to the troposphere. The step time for the chemical computation was changed from 15-30 minutes in the previous version to a few minutes in the present version, i.e., set to be one tenth of the dynamical time step; (2) As a meteorological data assimilation in the model, a nudging method was applied, where zonal and meridional wind velocities and temperatures obtained from observations were input into the CTM at six-hourly time intervals, as detailed in the next paragraph below.

The model values of zonal and meridional wind velocities and temperature were nudged toward the values of the input data. The input data are European Centre for Medium-Range Weather Forecasts (ECMWF) operational data below 10 hPa. Above 10 hPa, where no ECMWF data exist, the monthly and zonal means of zonal wind and temperature data of COSPAR International Reference Atmosphere 86 (CIRA1986) (Rees et al., 1990) were used. The meridional wind above 10 hPa was not nudged but calculated by the continuity equation in the model. This simple meteorological data assimilation works quite well in the model to simulate day-to-day variations in O$_3$, N$_2$O, and CH$_4$ (Akiyoshi et al., 2002a, b). The nudging time scale was set to be 1 day. The outputs from the CTM were archived at the interval of 1 day at 00UT. O$_3$ chemical forcing terms were output as daily averaged values.

3. Ozone Simulation for Recent Years

Figure 1 shows total ozone variations simulated by the CCSR/NIES nudging CTM and observed by Earth Probe Total Ozone Mapping Spectrometer (EP-TOMS) at Kiruna, Sweden (67.4°N, 20.2°E), at Sapporo, Japan (43.1°N, 141.3°E), and at Tsukuba, Japan (36.1°N, 140.1°E) for 1997 and 2000. The model simulates the observation quite well in terms of the absolute values as well as seasonal variation. The model also simulates the difference in ozone variation between 1997 and 2000: The larger ozone fluctuations at Kiruna and the larger ozone values at Sapporo in 2000 than in 1997 resulted from the fact that the Arctic polar vortex was more unstable in 2000 than in 1997.
Figure 1 Variations in total ozone amount at Kiruna (67.4°N, 20.2°E, top), Sapporo (43.1°N, 141.3°E, middle), Tsukuba (36.1°N, 140.1°E) for 1997 and 2000. The blue line indicates CTM results and the pink line indicates the EP-TOMS observation. The horizontal axis represents the day number from January 1.

4. Numerical Experiments and Results

Three numerical experiments were performed to study the effects of bromine species on ozone destruction. The experiments were summarized in the Table 1. The first experiment is a control run, where both bromine species and heterogeneous reactions were included. In the second experiment, all the heterogeneous reactions were ignored, but gas phase reactions of bromine species were included. In the third experiment, bromine species and the heterogeneous reactions related to bromine species were ignored, but the heterogeneous reactions related to chlorine species, N₂O₅, and HNO₅ were included. The difference between
the first experiment and the second indicates the effect of heterogeneous reactions. The difference between the first and the third indicates the effect of bromine species.

Table 1  Design for numerical experiment.

<table>
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4.1 Effects of Bromine Species on Ozone Destruction through Heterogeneous Reactions

Figure 2 shows the calculated seasonal variation of the zonal-mean mixing ratio of bromine species at 80.3°N in 1997. The left panel represents the result from the control run (Experiment 1). The right panel represents the result from the Experiment 2. The important heterogeneous reactions for bromine species are \( \text{BrONO}_2 + \text{H}_2\text{O} \rightarrow \text{HOB} + \text{HNO}_3 \) and \( \text{HOB} + \text{HCl} \rightarrow \text{BrCl} + \text{H}_2\text{O} \). Without heterogeneous reactions, major component of bromine species during the Arctic winter is \( \text{BrONO}_2 \), while \( \text{BrCl} \) is the major component when heterogeneous reactions work. \( \text{BrONO}_2 \) is converted to \( \text{BrCl} \) effectively on the liquid Polar Stratospheric Clouds (PSCs) during wintertime. Since \( \text{BrCl} \) is photolized by sunlight more easily than \( \text{BrONO}_2 \), those heterogeneous reactions made the rapid increase in the BrO concentration in the spring (See the BrO increase around Day 100). The rapid BrO increase before the ClO decrease in the spring enables the BrO-ClO catalytic cycle for ozone destruction to work effectively.

![Figure 2](image)

Figure 2  Seasonal variations in the zonal-mean mixing ratio (pptv) of bromine species at 80.3°N, 450 K in 1997. Result from the EXP 1 (the control run, left panel) and result from the EXP 2 (without heterogeneous reactions, right panel). The horizontal axis represents the day number from January 1. The variations of ClO and Cl\(_2\)O\(_2\) are also shown with the scale multiplied by 0.01.

4.2 Effects of Bromine Species on Ozone Destruction through Inter-annual Variation in Arctic Polar Vortex

In this section, we show the inter-annual variation in ozone destruction due to bromine species. The four cold Arctic winter years, 1994/95, 1995/96, 1996/97, and 1999/2000 were selected to show the inter-annual variation. In these winters, the minimum temperature at 50 hPa in the northern high latitudes was well below the NAT formation temperature (195 K)
and sometimes below the ice frost point (188 K) (Pawson and Naujokat, 1999; Manney and Sabutis, 2000; Sinnhuber et al., 2000). However, the seasonal evolution of the minimum temperature and the vortex stability are different among these years. The cold temperature below the NAT formation temperature appeared during December-January in 1994/95, mid-December-January-February in 1995/96, and mid-January-February-March in 1996/97. The temperature evolution in the winter of 1999/2000 is similar to that of 1995/96 winter, and the Arctic vortex conditions are also similar in January and February for these two years. But the Arctic vortex splitted into two parts in March, 2000. The more active vortex in March 2000 made a large ozone loss in the polar region in the same year (Millard et al., 2003).

Figure 3 shows the calculated seasonal variations of the zonal-mean mixing ratio of bromine species at 450 K, 80.3°N in 1995, 1996, 1997, and 2000. In the lower stratosphere at this latitude in 1997, a large spring maximum of BrO concentration was calculated (See around Day 100). Comparison of bromine species variations among these 4 years clearly shows that the distinct 1997 spring maximum in BrO concentration resulted from the buildup of Bry amount, which was caused by a stronger downward motion in the Arctic region in 1997 than in the other years, because Bry mixing ratio has a positive vertical gradient in the lower stratosphere. The stronger downward motion in the Arctic in 1997 is related to the unusually stable Arctic polar vortex in this year.

Figure 3  Seasonal variations in the zonal-mean mixing ratio (pptv) of bromine species at 80.3°N, 450 K in 1995, 1996, 1997, and 2000. The results from EXP 1 (the control run). The horizontal axis represents the day number from January 1. The variations of ClO and ClO2 are also shown with the scale multiplied by 0.01.
4.3 Effects of Bromine Species on Ozone Destruction through a Coupling with Chlorine Chemical Reactions

Figure 4 shows the seasonal variations in the zonal-mean chemical ozone destruction forcing of several catalytic cycles at 80.3°N, 450 K in 1997. The left figure shows the result from the EXP 1 (the control run) and the right figure shows the result from the EXP 3 (without bromine species). The figure suggests that the ClO-ClO ozone destruction catalytic cycles indicated by the red line and labeled by “[ClO₂]” was strengthen with bromine species. This is because BrCl was formed in the control run, and hence, more chlorine was activated from the reservoir species.

![Figure 4](image)

**Figure 4** Seasonal variation in the zonal-mean chemical ozone destruction forcing of the catalytic cycles (ppbv/day) at 80.3°N, 450 K in 1997. The results from EXP 1 (the control run, left), and the result from EXP 3 (without bromine species, right). The horizontal axis represents the day number from January 1.

5. Concluding Remarks

A new version of the CCSR/NIES nudging CTM was developed. Bromine chemical reactions and sulfate reactions were introduced in the model. The model simulates seasonal variation and inter-annual variation of total ozone amount very well. Bromine effects on ozone destruction were studied by performing a control model run and another two model runs where bromine species or heterogeneous reactions were ignored. Comparing the results from the control run with the other runs, we conclude that (1) heterogeneous reactions made the fast BrO increase in spring, which promotes the ozone destruction through the BrO-ClO catalytic cycle (2) unusually stable Arctic polar vortex in 1997 caused the build-up of BrO at high latitudes in the spring and increased BrO amount. (3) bromine species enhanced ClO-ClO ozone destruction cycle through the formation of BrCl in wintertime.
References


**Presentations and Publications**

**Original Paper:**

**Presentations:**

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A Study on the Polar Ocean Freshwater Budget and the Global Thermohaline Circulation Using the CCSR Sea Ice-Ocean Model (Iced-COCO)

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Abstract
The role of the freshwater/salinity budget of the northern hemisphere polar oceans in forming the global thermohaline circulation is investigated by use of a sea ice-ocean coupled general circulation model. Two aspects of the freshwater budget are focused on: one is the uncertainty in climatological sea surface freshwater flux datasets, and the other is the Arctic pathway of the fresh Pacific water. In the study of the uncertainty in climatological datasets, high sensitivity of the model-simulated Atlantic meridional overturning circulation to the river runoff into the Arctic Ocean is emphasized. As for the Arctic pathway of the Pacific water, a stabilizing effect of the pathway through the Canadian Archipelago is suggested, contrary to a previous study.

Keywords: OGCM, Thermohaline circulation, Freshwater budget, Atlantic meridional overturning circulation

1. Introduction

Stability and variability of the oceanic global thermohaline circulation is one of the focal points in discussing various kinds of climatic phenomena and changes, and its realistic reproduction is an essential part of the modeling efforts of the climate. The Atlantic meridional overturning circulation (AMOC), which is the most important part of the global thermohaline circulation, is believed to be driven by the deep convection taking place in the Labrador Sea and the Greenland Sea, and such deep convection is known to be susceptible to slight changes of salinity environment of the basins. Salinity of these polar seas is controlled by various factors: sea surface freshwater forcing, sea ice formation/melt and transport, and water exchange with other basins through narrow passages. For example, chemical tracer analyses show that the deep water formation in the Greenland Sea was significantly reduced since 1980 (Bönisch and Schlosser, 1995). The inactivation of the deep water formation is considered to be related to the advection of the low salinity water, known as the Great Salinity Anomaly (Dickson et al., 1988), which is closely linked to the sea ice transport through the Fram Strait (Håkkinen, 1993).

Both the Greenland Sea and the Labrador Sea are under the strong influence of the Arctic Ocean. The Arctic Ocean receives about 0.1 Sv (1 Sv = 1×10^6 m³/s) of freshwater as river runoff, which is the largest freshening source of the basin. The Bering Strait throughflow transports the water whose salinity is significantly lower than the mean salinity of the Arctic Ocean. The equivalent freshwater transport by the Bering Strait throughflow is about 0.05 Sv. These freshwater inputs into the Arctic Ocean eventually exit through the Canadian Archipelago and the Fram Strait into the Labrador Sea and the Greenland Sea, respectively, and thus influence the salinity environment of these basins.

This document deals with two aspects of the role of the northern high latitudes freshwater budget in the AMOC. The first part discusses the differences in widely used climatological
datasets for sea surface freshwater flux and their significance in simulating the AMOC by ocean general circulation models (OGCMs). The second part discusses the influence of the relatively low salinity Pacific water on the AMOC by focusing on its pathway.

2. Model and Forcing

2.1 Ocean Model

The OGCM used herein is Center for Climate System Research (CCSR) Ocean Component Model (COCO) version 3. The model explicitly predicts changes in sea surface elevation by the method of Killworth et al. (1991). The model vertical coordinate system is a hybrid of $\sigma$ (normalized depth) and $z$ (geometrical depth): the former is applied between the free surface and a fixed level in the upper ocean, and the latter below. The partial step formulation (Adcroft et al., 1997) is adopted for bottom topography representation. The model is horizontally formulated on the spherical coordinate system. Since the model adopted herein is global, the spherical coordinate system is rotated so that the north pole of the coordinate is on Greenland to avoid the coordinate convergence in the Arctic Ocean.

For the tracer equations, the uniformly third-order polynomial interpolation algorithm (Leonard et al., 1993), the harmonic isopycnal diffusion of tracer (Cox, 1987) and layer thickness (Gent et al., 1995), and harmonic vertical diffusion are used. The momentum equations are solved by the classical centered-in-space, centered-in-time algorithm, which guarantees the conservation of momentum and kinetic energy, together with the harmonic viscosity. The vertical diffusivity is estimated by the level two turbulence closure of Mellor and Yamada (1982). A background profile is prescribed for the vertical diffusivity to reproduce the Pacific deep meridional overturning circulation with realistic intensity (Tsuchino et al., 2000). It takes $1 \times 10^{-5}$ m$^2$/s at the surface and $3 \times 10^{-4}$ m$^2$/s at the deepest level of the model, with a sharp increase between 1000 m and 3000 m. The model incorporates a bottom boundary layer parameterization of Nakano and Suginoahara (2002).

Other model parameters and resolution are not common in the following two parts, so they are described in the relevant sections.

2.2 Sea Ice Model

The thermodynamic part of the sea ice model is the simplest zero layer model (Semtner, 1976), where the heat content of sea ice is neglected and only the latent heat of melt is taken into account for heat budget calculation. Sea ice exists only when and where the temperature of the top level of the ocean model is at the freezing point, and ice bottom temperature is given by that. The vertical heat flux through sea ice and air-ice interface depends on the temperature at the top of sea ice (or snow cover), and it is determined by the balance of them. If the diagnosed temperature goes beyond the melting point, it is reset to the melting point and the resulting heat imbalance is consumed to melt sea ice.

The dynamic part consists of advection and mechanical change of sea ice thickness, the latter of which represents the pressure ridge formation process due to sea ice convergence. In representing the mechanical characteristics, sea ice is treated as a continuum and a certain rheology is assumed to formulate its response to deformational forcing. The rheology adopted here is the elastic-viscous-plastic formulation of Hunke and Dukowicz (1997).

In sea ice thickness representation, the model adopts the two-category formulation, where concentration (fractional coverage) and mean thickness of the ice-covered region are
predicted in each grid. Sea ice is assumed to contain salinity of a constant value, which is taken to be 5 psu in this study.

2.3 Surface Forcing

The ice-ocean model is forced at the sea surface by heat, freshwater and momentum fluxes. The freshwater flux consists of evaporation, precipitation, and river runoff. The traditional restoring of sea surface salinity to climatology is not applied herein. Although this restore-free surface salinity boundary condition results in biases, some of which are significant, of salinity distribution, this choice is crucial in sincerely discussing the role of freshwater budget in the deep water formation and the consequent thermohaline circulation. The heat flux is derived from the surface radiative fluxes and the surface air temperature, humidity, and wind. The momentum flux is given by the two components of the sea surface wind stress. All the forcing data are taken from the Ocean Model Intercomparison Project (OMIP) forcing dataset (Rösk, 2001), which is daily climatology based on European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis, unless otherwise noted. The bulk formulae of Kara et al. (2000) are used in deriving the surface heat flux, which are also used by Rösk (2001) to close annual heat budget in compiling the climatological dataset.

3. Difference in Freshwater Flux Climatology Datasets and Its Influence on the AMOC

In this section, we show how large the impact of the difference in two commonly used climatological sea surface freshwater flux datasets on the AMOC is. More detailed description of the model and the results can be found in Oka and Hasumi (2003a).

3.1 Model Setup

The horizontal resolution of the model is 2°, and there are 42 levels in the vertical, five of which are in the upper 50 m and are formulated on the σ-coordinate. The latitudinal (in the rotated coordinate system) resolution is raised to 0.5° at northern high latitudes to improve the reproduction of the intricate current system in and around the Greenland-Iceland-Norwegian (GIN) Seas. This treatment is crucial for the reproduction of the AMOC in a coarse resolution OGCM (Oka and Hasumi, 2003b). The model is initiated by the Polar Science Center Hydrographic Climatology (PHC; Steele et al., 2001) and is integrated for 5000 years in each case. The state averaged for the last 100 years is used for the analysis.

3.2 Freshwater Flux Datasets

Two datasets for sea surface freshwater flux are used here. One is the OMIP forcing dataset described in the previous section, which provides all necessary forcing components (evaporation, precipitation and river runoff) in a consistent manner. The other consists of evaporation and precipitation climatology based on the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al., 1996) and river runoff climatology of Perry et al. (1996), the latter of which is the compilation of direct observations. This set of data is called NCEP forcing dataset henceforth. Note that river runoff of the OMIP dataset is derived from evaporation and precipitation on land by prescribing a catchment area for each of the represented rivers. There is no other
easily available global river runoff dataset. As for evaporation and precipitation, there are some other climatological datasets, mostly based on direct observations. However, we decided to use only reanalysis datasets, since it can provide evaporation and precipitation at the same time without data missing regions.

The annual-mean, zonally averaged profiles of evaporation and precipitation for the OMIP and NCEP datasets are illustrated in Figure 1. The difference in precipitation is significant especially in the tropics. Representation of tropical rainfall depends heavily on the parameterization of cumulus convection, and slight difference in this respect or in resolution may lead to such a large difference. As for evaporation, overall difference between the two datasets is relatively small.

![Figure 1](image)

**Figure 1** Annual-mean, zonally averaged profile of evaporation (left) and precipitation (right panel) in cm/yr. The solid line is for the OMIP dataset and the dashed line is for the NCEP dataset.

Annual-mean river runoff integrated over major basins is compared in Table 1 between the two datasets. The NCEP dataset gives smaller values, which may be attributed to the fact that the compilation of Perry et al. (1996) is limited to 981 major rivers.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Arctic</th>
<th>GIN Seas</th>
<th>Atlantic</th>
<th>Pacific</th>
<th>Indian</th>
</tr>
</thead>
<tbody>
<tr>
<td>OMIP</td>
<td>0.094</td>
<td>0.056</td>
<td>0.642</td>
<td>0.359</td>
<td>0.196</td>
</tr>
<tr>
<td>Perry et al.</td>
<td>0.075</td>
<td>0.033</td>
<td>0.425</td>
<td>0.179</td>
<td>0.092</td>
</tr>
</tbody>
</table>

**3.3 Sensitivity of the AMOC to Difference in Freshwater Flux Forcing Datasets**

The volume transport stream function of the annual-mean, zonally integrated meridional overturning circulation in the Atlantic is shown in Figure 2. The deep maximum at the equator is 12.3 Sv for the NCEP forcing case and is 6.3 Sv for the OMIP forcing case. Although the NCEP forcing case gives a value closer to observational estimates (e.g., 14 Sv by Schmitz, 1995), it does not necessarily mean that the NCEP forcing dataset is better than the OMIP forcing dataset. It might be simply the case that the current model is better tuned for the NCEP forcing dataset. So, we do not discuss which dataset is better or worse. We just discuss
here which aspect of the difference between the two forcing datasets significantly affects the AMOC.

Figure 2  Stream function of the annual-mean, zonally integrated volume transport in the Atlantic for the cases forced by the OMIP (left) and NCEP (right) dataset. Contour interval is 1 Sv.

A number of experiments are carried out with replacing a part of the NCEP forcing freshwater flux components by the corresponding part of the OMIP forcing dataset. The replaced component and/or area in each case is as follows: the whole river runoff (R-GLB); river runoff into the Arctic Ocean (R-ARC); river runoff into the GIN Seas (R-GIN); river runoff into the North Atlantic to the north of 60°N (R-ATL); evaporation and precipitation in the North Atlantic between 30°N and 60°N (EP-NM); difference in evaporation minus precipitation between the Atlantic and the Pacific at low latitudes (EP-L); difference in evaporation minus precipitation between the latitudinal bands 30°N-30°S and 30°S-90°S (EP-S). The results are summarized in Table 2.

Table 2  Sensitivity of the deep maximum of the AMOC at the equator to changes in the freshwater forcing dataset. ΔF is the modified amount of the freshwater flux (in Sv), and ΔV is the consequent change in the AMOC transport at the equator (in Sv).

| Experiment | ΔF  | ΔV  | |ΔV/ΔF|
|------------|-----|-----|------|
| R-GLB      | +0.019 | -5.07 |     |
| R-ARC      | +0.023 | -0.84 | 45.16 |
| R-GIN      | +0.017 | -0.86 | 37.23 |
| R-ATL      | -0.065 | -0.57 | 34.13 |
| EP-NM      | +0.310 | -0.72 | 11.01 |
| EP-L       | +0.450 | +3.48 | 11.23 |
| EP-S       |     | -2.08 | 4.62 |

The reduction of the AMOC in R-GLB is comparable to that obtained by replacing the entire of the freshwater flux dataset, and more than 40% of that reduction can be attributed to the river runoff difference at northern high latitudes (R-ARC, R-GIN, and R-ATL). The northern high latitudes river runoff accounts for only 13-14% of the globally integrated value (0.117 of 0.804 Sv in the NCEP dataset and 0.178 of 1.347 Sv in the OMIP dataset), but the difference in it between the two datasets has a large impact on the AMOC. Although the
influence of the difference in freshwater forcing on the AMOC is not a simple linear combination of the contribution from river runoff and evaporation minus precipitation, the major contributor in the difference in the AMOC between the two forcing datasets is river runoff, especially at northern high latitudes. However, when we further decompose evaporation minus precipitation flux by considering the background processes, the differences in it between the two datasets have larger impacts on the AMOC.

The basin-averaged freshwater loss from the sea surface is positive for the Atlantic and is negative for the Pacific, and a significant part of the imbalance is compensated by the atmospheric freshwater transport from the Atlantic to the Pacific, primarily across Central America. A change of the order of 0.1 Sv in this atmospheric interbasin freshwater transport has a significant influence on the AMOC (Hasumi, 2002a). The experiment EP-L is intended to assess the impact of the difference in the freshwater forcing datasets in this respect. The Atlantic-Pacific contrast of the evaporation minus precipitation flux at low latitudes (30°N-30°S) differs by 0.31 Sv between the two forcing datasets. When the contrast in the NCEP forcing is raised by increasing freshwater loss (and gain) in the Atlantic (and the Pacific) to the value of the OMIP forcing, the AMOC is intensified by 3.5 Sv, which is significant enough as compared with the impact of the difference in river runoff.

With regard to the energy budget of the thermohaline circulation, the AMOC is driven primarily by the buoyancy loss at northern high latitudes and the buoyancy gain in the Southern Ocean (Hasumi and Sugino, 1999). Poleward atmospheric freshwater transport in the Southern Hemisphere affects the latter part of the buoyancy budget, and its effect is examined by the experiment EP-S. As seen in Table 2, the replacement results in reduction of the AMOC intensity by 2 Sv. The effects assessed by these two experiments are not independent, but it is certain that the two effects are canceling out each other to reduce the impact of the entire difference in the evaporation minus precipitation flux on the AMOC.

4. Arctic Pathway of the Pacific Water and Its Influence on the AMOC

Influence of the relatively fresh Pacific water into the Arctic Ocean on the AMOC is discussed in this section. More detailed description of the model and the results can be found in Komuro and Hasumi (2003).

4.1 Model Setup

The horizontal resolution is 1° and there are 40 vertical levels, five of which are in the upper 50 m and are formulated on the σ-coordinate. In the forcing dataset, river runoff into the Arctic Ocean is artificially reduced by 30 % in order to secure a good control case result. Without such modification, runoff of the OMIP dataset would lead to very weak AMOC as seen in the previous section. Note that the Arctic river runoff of Perry et al. (1996) is 20 % smaller than that of the OMIP dataset, so the current artificial modification is not too large compared with the difference in the climatological datasets; though the result of the previous section also indicates even a slight change in the Arctic river runoff has a very large impact on the AMOC.

There are two cases of experiments: the water pathway through the Canadian Archipelago is open (as is in reality) in one case, and it is closed in the other case. The former is referred to as OPEN and the latter CLOSED. This water pathway is not naturally resolved at 1° horizontal resolution, so it is artificially widened in OPEN. The model is initiated by the PHC
climatology, and the integration is continued for 1500 years until a virtually steady state is obtained. The state averaged over the last five years is used for analysis.

4.2 Results

The annual-mean northward volume transport through the Bering Strait for OPEN is 1.7 Sv, which is somewhat larger than observed values, and the volume transport out of the Arctic Ocean through the Canadian Archipelago is 1.2 Sv. There are two exit routes for the Pacific water out of the Arctic Ocean: the Canadian Archipelago and the Fram Strait. A model experiment where a virtual trace source is placed at the Bering Strait shows that the primary exit route is the Canadian Archipelago. This is consistent with observational knowledge (Jones et al., 2003).

The volume transport stream function of the annual-mean AMOC is shown in Figure 3. The deep maximum at the equator is 12.4 Sv for OPEN and is 8.7 Sv for CLOSED. In both the cases, deep convection (deep water formation) is taking place in the Greenland Sea and the Labrador Sea as in reality, although the latter is occurring not in the central part but at the mouth of it. Decomposition of the AMOC volume transport into the contribution from each of the deep water formation is possible when the stream function is drawn with choosing density as the vertical coordinate (Figure 4): the deep maximum at about 60°N represents the contribution from the Greenland Sea, and the rest is attributable to that from the Labrador Sea. In this view, the both of the contribution are intensified, or the both of the deep water formation are activated, by opening the Canadian Archipelago water pathway.

![Figure 3 Stream function of the annual-mean, zonally integrated volume transport in the Atlantic for OPEN (left) and CLOSED (right). Contour interval is 2 Sv.](image)

One might expect that opening the Canadian Archipelago pathway would activate deep water formation in the Greenland Sea but inactivate that in the Labrador Sea, as the freshening effect of the Pacific water is reduced for the Greenland Sea, while it is enhanced for the Labrador Sea. Indeed, such a result was obtained in a previous study of this kind (Wadley and Bigg, 2002), and the total AMOC volume transport was reduced by opening the pathway in that study.

What's happening in the Greenland Sea is consistent with that intuitive conjecture and is common between the previous and current studies. As a result of the reduction of fresh water inflow into the Greenland Sea, density of the deep water formed there is raised by about 0.2 in
\(\sigma_2\) (Figure 4) by opening the Canadian Archipelago, and the transport of southward overflow across the sill at about 60°N is increased by about 2 Sv.

![Figure 4](image_url)  

**Figure 4**  Same as in Figure 3 except that the vertical coordinate is \(\sigma_2\). Contour interval is 1 Sv.

In the Labrador Sea of the present model, on the other hand, the water coming through the Canadian Archipelago does not flow directly into the region of deep convection but is diverted by the cyclonic circulation around the deep convection site. Furthermore, the influence of raised salinity in the Greenland Sea is transported by the East Greenland Current and then the West Greenland Current into the region of deep convection in the Labrador Sea, thus activating the deep water formation there. The cause of the different response of the Labrador Sea deep water formation to the opening of the Canadian Archipelago between the previous and current studies may be in the resolution of the basin. In the previous study, the Labrador Sea was not sufficiently resolved to reproduce the cyclonic circulation, which exists in reality, around the deep convection site.

Although the present model is not satisfactory very much in representing the deep water formation sites and the circulation in the surrounding regions, the basic processes and their geographical distribution are qualitatively captured, at least in a better form than in previous studies. So, we believe that the present model result is valid in assessing the effect of the Arctic pathway of the Pacific water on the AMOC. This study indicates the significance of the water pathway through the Canadian Archipelago in stable existence of the AMOC, together with the significance of the current system forming the cyclonic circulation around the Labrador Sea deep convection site.

5. Concluding Remarks

Restoration of sea surface salinity implies the flux of salt through the sea surface, which, of course, is physically unreasonable. Still, the methodology to reproduce the oceanic global thermohaline circulation by OGCMs without restoring sea surface salinity to observed values has not been well established, nor have such attempts been successful enough. The sources of the difficulty are categorized roughly into two: one is in the model physics, and the other is in the accuracy of freshwater forcing datasets. The "haline" aspect of the thermohaline circulation can never be faithfully examined when salinity restoration is adopted. Moreover, since the thermal and haline aspects are inseparable in the physics of the thermohaline circulation,
further progress of the study on the thermohaline circulation depends crucially on the solution of this problem.

As recognized in this document, our sea ice-ocean model succeeds in achieving that goal to some good degree, and a meaningful study on the thermohaline circulation from the viewpoint of freshwater budget is enabled. The achievement is brought about by many factors: incorporation of up-to-date numerical algorithms and physical parameterizations; good tuning of the model, especially in terms of the choice of mixing parameters; clarification of key processes and their good representation, including a suggestion for an appropriate choice of model resolution (Oka and Hasumi, 2003b). See also the review by Hasumi (2002b) for the factors to be considered with respect to modeling of the global thermohaline circulation.

Water cycle is one of the most important viewpoints describing the link of the elements composing the climate system. The ocean is the largest reservoir of water in the climate system, and its thermohaline circulation, which is the sole process linking the whole depths of the ocean, is highly dependent on the exchange of freshwater with other climatic subsystems. Understanding of the physics of the climate system is not realized without taking into account the freshwater aspect of the oceanic circulation, and the proper modeling of the climate system does rely on the proper representation of the oceanic freshwater/salinity budget, which is the very thing this study intends.

References


**Publications and Presentations**

**Original Papers and Reviews:**


**Conference Report:**

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Development and Application of Parallel Atmospheric Transport Model to Inverse Modeling of Global Carbon Cycle

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Abstract
Following an approach developed for Transcom-3 CO₂ inverse modeling experiment, we prepared a forward simulation based on interannually varying winds by NCEP reanalysis, aiming at global carbon cycle analysis during last 20 years and level-3 intercomparison study. NCEP pressure level data are used with diagnostically calculated upper level vertical winds. Several transport model modifications were introduced in order to achieve parallel calculation of multiple regional signals on computer clusters. An analysis of the seasonal and interannual variability of the land and atmospheric CO₂ fluxes at regional scale (64 regions) reveals connection between the atmospheric CO₂ variability and climate variability in the interannual/decadal time scale. Our results suggest that the inverse model fluxes for (mostly tropical) land and ocean regions are correlated with the Southern Oscillation Index (SOI).

Keywords: Atmospheric CO₂, Inverse model, Climate change

1. Objective

Atmospheric CO₂ concentration measurements provide a consistent record, containing large amount of information on the global carbon cycle, of both terrestrial and oceanic sources. Atmospheric transport models combined with observations are being used with some success in deciphering seasonal/interannual variability and spatial distribution of the surface to atmosphere fluxes of CO₂. The details and quality of the inverse atmospheric transport analysis depends critically on the spatial and temporal resolution of the forward transport simulation and inverse model retrievals of the surface fluxes. Focus of our work is to develop a technique of the inverse model analysis at high spatial and temporal resolution, based on a computationally efficient parallel atmospheric transport model optimized for clustered and vector supercomputers.

2. Materials and Methods

In the inverse modeling of the global carbon cycle, information on the CO₂ flux distribution is retrieved from atmospheric CO₂ observations and known information on spatial and temporal distribution of the surface CO₂ exchange (Enting et al., 1995, Bousquet et al., 2000). The accuracy of the inverse model retrieval is dependent on the transport model accuracy, the method of solution and a number of source regions and time intervals to analyze for. Later effect was described as aggregation error by Kaminski et al. (2001), concluding that
the number of source regions should as large as possible within the computing capability available. In our study, global atmospheric transport model (Maksyutov and Inoue, 2000) is used to simulate the monthly and yearly tracer pulse functions. Several modifications in the transport model were introduced in order to achieve parallel calculation of multiple regional signals on multiprocessor supercomputer systems with MPI interface. NIES supercomputer system was used for the model development and the production runs completed on the Earth Simulator.

The transport model was redesigned to run efficiently on multiple processors and was used with up to 72 processors on NEC supercomputers.

The transport of different groups of tracers is performed independently on each processor element, providing good scalability and performance. Two approaches to achieve scalability have been tested. In simple setup, each processor element reads the wind fields and does preprocessing. In more advanced version, one processor element does reading and preprocessing and others receive the preprocessed data via message passing mechanism. Both approaches lead to significant speedup in the calculation, delivering performance only possible on fast and massively parallel computers. In fact, the only other detailed inverse model study by Rodenbeck et al. (2003) was produced on similar NEC SX system at MPI Meteorology in Hamburg.

Following Transcom-3 interannual inverse model analysis technique, similar to Rayner et al. (1999), we prepared a forward simulation based on interannually varying winds by NCEP reanalysis, aiming at global carbon cycle analysis during last 20 years, then solved for monthly fluxes for each of the 64 (42 land and 22 ocean) inverse model regions. The transport model has 15 sigma levels in the vertical, and 2.5x2.5 horizontal resolution. Pressure level data by NCEP were used, and missing vertical velocities above 100 mbar were filled with ones following calculated isentropic trajectories.

The seasonal and interannual variability in the land-atmosphere and ocean-atmosphere fluxes is obtained by applying a Bayesian inverse model to retrieve monthly flux corrections constrained by atmospheric observations of CO₂ and prescribed patterns of the anthropogenic, terrestrial and oceanic fluxes.

Another problem we studied was the applicability of future satellite observations to inverse model analysis of the global carbon cycle (e.g., Rayner and O'Brien, 2001). The studies comparing efficiency of the ground-based and satellite observations in the CO₂ sources and sinks estimation with inverse models used limited number of source regions against large number of the remote sensing observation, and the results usually favor the higher accuracy ground based observations (Patra et al., 2003). Thus, it is essential to have a larger number and smaller sizes inverse model regions for analysis of the satellite data utility. In our setup, we studied the flux uncertainty reduction by inverse model for a grid of 10 by 15 degrees latitude–longitude regions, 432 regions in total. In the annual average model, the annual pulses from 432 regions are calculated and the inverse model uncertainty reduction is estimated. The details are reported in Maksyutov et al. (2003).

3. Results

An analysis of the seasonal and interannual variability of the land and atmospheric CO₂ fluxes at regional scale (64 regions) reveals connection between the atmospheric CO₂ variability and climate variability in the interannual/decadal time scale. Our results suggest that the inverse model fluxes for (mostly tropical) land and ocean regions are correlated with the Southern Oscillation Index (SOI).
Generally it can be seen that larger oceanic uptake (net sink of CO$_2$), and release from the land (net source) during the onset of an El-Nino event. The net oceanic sink is attributed to the less intense upwelling primarily in the equatorial Pacific region. The land sources are triggered by drier atmospheric conditions (drought) and enhanced forest fires during the El Niño period, e.g. the Indonesian forest fire in 1997-1998 (Fig. 1).

![Figure 1 Anomalies in the global total, land and ocean fluxes of CO$_2$ as estimated using a 64-region time-dependent inverse model and atmospheric CO$_2$ data from 87 stations. The southern oscillation index (SOI) is plotted above to show the co-variability.](image)

In the evaluation study of satellite data utilization, spatial distribution of the inverse model uncertainty reduction was estimated for 3 sets of projected observations from space: 1) column observations (in reflected sun light), 2) free tropospheric observations (above 700 mbar), and 3) column observations over ocean (in reflected sun light, in sun glint conditions). The information content from the column observations (Case 1) appears to be the largest and the corresponding maps of flux uncertainty reduction are presented in detail in a recent publication (Maksyutov et al., 2003).

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**References**


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Equilibrium Response of the Atlantic Thermohaline Circulation to Climate Change – Theory and CGCM Experiments –

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Abstract
To understand the behavior of the Atlantic thermohaline circulation (THC) to climate changes, a theoretical consideration based on the conceptual 4-box model has been performed. It is shown that the long-term response of the THC to climate change is determined by its regime. If the present-day THC is in the thermally driven regime, the long-term response is consistent with the transient response and the THC will be reduced in the case of global warming. However, if the current THC is in the thermohaline driven regime, the long-term response differs from the transient response and there exists a possibility that the THC will be enhanced in the distant future where global warming goes forward. The results of some general circulation model (GCM) simulations suggest that the current Atlantic THC is in the thermohaline driven regime. This consequence also explains the reason of the weakened THC in glacial maximum.

Keywords: Atlantic thermohaline circulation, Coupled GCM, Box model

1. Introduction

Paleoceanographic evidence suggests that the Atlantic thermohaline circulation (THC) in glacial periods gradually weakens with progress of the glacial cycle and reaches its minimum at the glacial maximum (e.g. Duplessy et al., 1988). Moreover, some evidence suggests the enhancement of the THC at the first stage of the glaciation (McManus, 2002). To understand such behaviors of the THC, it is important which regime the present-day THC is in. Some paleoceanographers think that the current THC is driven by the salinity contrast between the Atlantic and Pacific (Weyl, 1968; Broecker and Denton, 1989). This idea is close to the “haline driven regime” of the Stommel’s 2-box model of the THC (Stommel, 1961). It explains the weakening of the THC in glacial periods, but can not explain the initial enhancement of the THC at glacial inceptions.

On the other hand, the Intergovermental Panel on Climate Change (IPCC) (1996, 2001) reports that most climate models predicted a weakening of the Atlantic THC in global warming experiments where the atmospheric greenhouse gas concentrations were increased according to some scenarios. Such a THC weakening is often explained by the following two mechanisms. 1) The temperature rises caused by the global warming are larger in high latitudes. It reduces the atmospheric cooling of the ocean and inhibits the density increases of the surface sea-water at the deep-water formation regions. This causes the THC weakening. 2) Increases of water vapor due to atmospheric temperature rises activate the global hydrological cycle and increase freshwater supply at high latitudes. This decreases the density of surface sea-water at the deep-water formation regions. This also causes the THC weakening (Manabe et al., 1991). Behind above two mechanisms, there exists a view that the current Atlantic THC is thermally driven and inhibited by the freshwater forcing at high latitudes (e.g. Rahamstorf, 1996). This view of the THC comes from the “thermally driven regime” of the Stommel’s 2-box model (Stommel, 1961), and is often used in explaining the behavior of THC in GCMs (e.g. Bryan, 1986; Manabe and Stouffer, 1988; Manabe et al., 1991; Rahamstorf, 1996). However, such a view can not explain the weakening of the THC in glacial periods by itself. Moreover, in long-term experiments conducted by Manabe and Stouffer (1994), the THC, that has once weakened by the
global warming, recovers the original intensity hundreds or thousands years after and becomes stable in a slightly strengthened level. It is impossible to explain this result based on the view that the current THC is in the thermally driven regime.

To clarify such confusion, the authors have performed a theoretical consideration based on the Rahmstorf's 4-box model (Rahmstorf, 1996), and have conducted some general circulation model (GCM) experiments. These suggest that the current THC is in the “thermohaline driven regime” of the Stommel’s or Rahmstorf’s box model.

2. Theory

2.1 Equilibrium Solutions of the 4-box Model

Figure 1 shows the conceptual box model introduced by Rahmstorf (1996) to represent the Atlantic THC. This model has four boxes. We assume that box-2 represents the high latitude in the North Atlantic. But we make no assumption which regions the other boxes specifically represent. Each box, except box-4, receives thermal forcing due to environmental temperature $T_1^*$ and fresh water forcing due to environmental freshwater transport $F_a, F_b$. Here, the word environment mainly represent the atmosphere. But it also represents the synthetic effect from other ocean basins and/or continents surrounding the Atlantic. These external forcing make the temperature contrast and salinity contrast between box-1 and box-2. These cause the density contrast between box-1 and box-2 and it drives the Atlantic THC. This linkage is represented by the equation

$$q = k(\beta \Delta S - \alpha \Delta T),$$

where $\Delta T = T_2 - T_1$ and $\Delta S = S_2 - S_1$. $T_i$ and $S_i$ denote the temperature and salinity in box-i. $k$ is a coefficient associating density contrast to the volume transport of the THC. $\alpha$ and $\beta$ are coefficients for thermal expansion and saline contraction respectively. The specific expressions of $\Delta T$ and $\Delta S$ in equilibrium are given in appendix (equations (11) and (12)). By substituting (11) into (1) and using relation (12), we obtain a relation about the equilibrium solutions as:

$$q^2 = \begin{cases} -qk\alpha\Delta T + k\beta S_0 F_a, & \text{for } q > 0 \\ -qk\alpha\Delta T + k\beta S_0 F_b, & \text{for } q < 0 \end{cases}$$

where $S_0$ is a reference salinity, which convert water fluxes to salt fluxes, treated as a constant. Figure 2 shows the graph of above relation for solutions $q > 0$. The abscissa in Fig. 2 represents the environmental freshwater transport $F_a$ from box-3 to box-1 and the ordinate represents the THC intensity $q$. The shape of the graph is approximately parabolic and depends on the values of $T_1^*$ (graphs for 4-type of $\{T_1^*\}$ are shown in Fig. 2). Stable solutions illustrated in Fig. 2 are classified into three categories. We name these as follows.

1. thermally driven regime
   The second quadrant of Fig. 2 (upper branch). $q > 0$ and $F_a < 0$. $T_1 > T_2$ and $S_1 > S_2$. The coldness of the box-2 dominates the saltiness of the box-1 in density contrast and it drives the circulation. The temperature contrast strengthens the circulation, while the salinity contrast inhibits it.

2. thermohaline driven regime
   The first quadrant of Fig. 2 (upper branch). $q > 0$ and $F_a > 0$. $T_1 > T_2$ and $S_1 < S_2$. The coldness and the saltiness of the box-2 drive the circulation. Both the temperature contrast and the salinity contrast strengthen the circulation.
3. haline driven regime
The first quadrants of Fig. 2 (lower branch), \( q > 0 \). \( T_1 < T_2 \) and \( S_1 < S_2 \). The saltness of the box-2 dominates the coldness of the box-1 in density contrast and it drives the circulation. The salinity contrast strengthens the circulation, while the temperature contrast inhibits it.

2.2 Equilibrium Response of the THC in the Box Model
As the global warming proceeds, the temperature difference between the high latitude and the low latitude decreases. It causes a decrease of the absolute values of the net heat flux from the atmosphere to the ocean. In the 4-box model, this process is essentially represented by a decreasing of the environmental temperature differences \( |\Delta T^*| = |T_2^* - T_1^*| \) and \( |\Delta T^*| = |T_3^* - T_1^*| \) where the order relation of \( T_i^* \) is invariant (by expression \( |\Delta T^*| \to 0 \) we symbolically represent such changes). An enhancement of the environmental fresh water transport due to global warming is represented by an increase of the absolute values of \( F_a \) and/or \( F_b \). Therefore, we can separate the atmospheric forcing to the ocean under global warming into two parts, and represent those in the 4-box model by changing \( |\Delta T^*| \) and \( |F_a| \) independently or synchronously. In this subsection, we examine the equilibrium responses of \( q_i \), \( \Delta S \) and \( \Delta T \) to such changes in external forcing.

For later use, we examine some relations for equilibrium solutions. From equation (12), we
can obtain a relation
\[ \Delta S = \frac{S_0}{q} F_a \, . \]  
(3)

This relation holds in any case.

Next, we examine relations those hold in cases of \( \Delta T < 0 \) and \( q > 0 \). From equation (11), in this case, we can see that the equilibrium temperature difference \( \Delta T \) is a monotonously increasing function of \( q \) (or \( q/\gamma \)) for \( q > 0 \) (\( \gamma \) represents the relaxation time of the system; see appendix). Therefore, an equation

\[ k\alpha \Delta T + q = 0 \]

has only one solution for \( q > 0 \). We put the solution as \( q_0 \). \( q_0 \) is the value of \( q \) which distinguishes the thermohaline driven regime and the thermally driven regime. The point \( (0, q_0) \) represents the intersection point of the parabolic curve (2) and ordinate axis. Detailed expression of \( q_0 \) is given in equation (13). Since \( q_0 \) decreases monotonically following a decrease of \( \Delta T \), the intersection point \( (0, q_0) \) shifts downward by a decrease of \( \Delta T \) (Fig. 2). By using \( q_0 \), we can approximate \( \Delta T(q) \) and \( \Delta S(q) \) as

\[ \Delta T(q) \approx -\frac{q_0}{k\alpha} + \frac{A}{k\alpha} (q - q_0) \]

(4)

\[ \Delta S(q) \approx \frac{B}{k\beta} (q - q_0) \]

(5)

where

\[ A = \frac{k\alpha (\Delta T^* + \Delta \tilde{T}^*)}{\gamma} \left( 1 + 3 \left( \frac{\beta}{k\alpha} \right)^2 \right) \]

\[ B = 1 + A \, . \]

Since \( \Delta T \) is a monotonic increasing function of \( q \), we can conclude \( A > 0 \) and \( B > 0 \). The property \( B > 0 \) is also followed by the fact that the sign of \( \Delta S \) must be consistent with the sign of \( q - q_0 \) on either side of \( q_0 \). Therefore, the function \( \Delta S(q) \) is an increasing function of \( q \).

Furthermore, we examine the case \( \Delta T > 0 \) and \( q > 0 \). In this case, we can approximate \( \Delta T(q) \) and \( \Delta S(q) \) in the neighborhood of \( q = 0 \) \((q > 0)\) as

\[ \Delta T(q) \approx \Delta T^* - (\Delta \tilde{T}^* - 2\Delta T^*) \frac{q}{\gamma} \]

(6)

\[ \Delta S(q) \approx \frac{\alpha}{\beta} \Delta T^* + \frac{1}{k\beta} \left( 1 - \frac{k\alpha}{\gamma} (\Delta \tilde{T}^* - 2\Delta T^*) \right) q \, . \]

(7)

The behavior of \( \Delta T \) depends on the order relation of \( \Delta T^* \) and \( \Delta \tilde{T}^* \). If \( \Delta \tilde{T}^* > 2\Delta T^* \), then \( \Delta T(q) \) is a monotonously decreasing function. If \( \Delta \tilde{T}^* < 2\Delta T^* \), then \( \Delta T \) increases with a increase of \( q \). (But this increase in \( \Delta T \) is restricted for small \( q \). \( \Delta T \) is a decreasing function for large \( q \).) However, \( \Delta S(q) \) is an increasing function if \( \frac{k\alpha}{\gamma} \) is small enough. From equation (7), we can see that \( \Delta S \) tends to the value \( \frac{\beta}{\alpha} \Delta T^* \) if \( q \) tends to 0. Therefore, from relation (3), we can express \( q \) as

\[ q \approx \frac{\beta}{\alpha} \frac{S_0}{\Delta T^*} F_a \, . \]

(8)

This relation shows that the THC intensity \( q \) is proportional to \( F_a \) and inversely proportional to \( \Delta T^* \). Therefore, the haline branch of the diagram shifts upward with a decrease of \( |\Delta T^*| \).

From these relations and Fig. 2, we can mathematically analyze the equilibrium response of the THC in the 4-box model to the changes in external forcing. The results are summarized in Table 1 and 2.
Table 1  Equilibrium responses of \( q \) and \( \Delta S \) in the 4-box model to changes in external forcing. The case \( \Delta T < 0 \).

<table>
<thead>
<tr>
<th>Forcing</th>
<th>Thermal ( F_a &lt; 0 )</th>
<th>Thermohaline ( F_a &gt; 0 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( F_a )</td>
<td>( \Delta T^* )</td>
<td>( q \Delta S(\cdot) \Delta T(\cdot) )</td>
</tr>
<tr>
<td>( \sqrt{\cdot} )</td>
<td>( - )</td>
<td>( \sqrt{\cdot} )</td>
</tr>
<tr>
<td>( \sqrt{\cdot} )</td>
<td>( \sqrt{\cdot} )</td>
<td>( \sqrt{\cdot} )</td>
</tr>
<tr>
<td>( \sqrt{\cdot} )</td>
<td>( \sqrt{\cdot} )</td>
<td>( \sqrt{\cdot} )</td>
</tr>
</tbody>
</table>

Table 2  Equilibrium responses of \( q \) and \( \Delta S \) in the 4-box model to changes in external forcing. The case \( \Delta T > 0 \) (haline driven).

<table>
<thead>
<tr>
<th>Forcing</th>
<th>Haline ( (\Delta T^* &gt; 2\Delta T^*) )</th>
<th>Haline ( (\Delta T^* &lt; 2\Delta T^*) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( F_a )</td>
<td>( \Delta T^* )</td>
<td>( q \Delta S(+) \Delta T(+) )</td>
</tr>
<tr>
<td>( \sqrt{\cdot} )</td>
<td>( - )</td>
<td>( \sqrt{\cdot} )</td>
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</tbody>
</table>

3. Coupled GCM Experiments

To investigate the regime of the Atlantic THC, the authors conduct a set of CGCM experiments. The model used here is the MRI-CGCM2.3 which has T42L30 AGCM and 144x111x23 grid OGCM with flux adjustment (a detailed model description is in Yukimoto et al., 2001). The experimental design is as follows.

1. Conduct two base-line integrations for 140 model years. One is a control integration conducted under the present-day greenhouse gas (GHG) concentrations. The other is a transient experiment where the GHG concentrations are increased by 1% per year. And save the daily water fluxes (precipitation minus evaporation plus river-runoff; P-E+R) from AGCM to OGCM.

2. Conduct two more (partially coupled) integrations for 140 model years. One is conducted under the present-day GHG concentrations but the daily surface water flux given to OGCM is taken from the saved data generated by the transient base-line run. The other is conducted under a similar condition to the transient base-line run but the daily surface water flux given to OGCM is taken from the saved data generated by the control base-line run.

Thus the four experimental integrations are conducted. We name those ctrl, tran, RcWt and RtWc, respectively.

The idea of these partially coupled experiments is due to Mikolajewicz and Voss (2000) and Dixon et al. (1999). They used such experiments to investigate the dominant forcing (thermal of freshwater) in weakening the THC under global warming. This idea is quite suitable for our purpose.

Figure 3 shows the timeseries of the THC intensity index in each runs. The THC intensity index is taken from the value of the Atlantic overturning stream function at the point of 30°N and 1500 m depth. Table 3 shows the trend of the changes in salinity contrast and temperature contrast between the northern high latitude and the low latitude of the Atlantic Ocean during the integration period.
Figure 3  Timeseries of THC intensity index in water flux exchange runs with MRI-CGCM2. THC index is the value of the Atlantic overturning stream function at the point of 30S, 1100m. 9-year running mean is applied.

| Table 3  Results of CGCM experiments. |
|---------|----|---|---|-------------------|
| run     | q  | S₂ | T₂ | (forcing)        |
| RcWt    | ✓  | ✓  | ✓  | (freshwater)     |
| RtWc    | ✓  | ✓  | ✓  | (thermal)        |
| tran    | ✓  | ✓  | ✓  | (freshwater + thermal) |

Table 3 clearly shows that both the salinity contrast and the temperature contrast between the high latitude and the low latitude increase in the latter half of the integration. By comparing this result with the theoretical consequence obtained in previous section (Table 1 and Table 2), we can conclude that the THC in MRI-CGCM is in the thermohaline driven regime.

4. Summary

To interpret the paleoclimatic data and results of GCM simulations for climate changes, a theoretical consideration based on the Rahmstorf’s 4-box model has been performed. The equilibrium responses of the THC to the freshwater forcing and thermal forcing are mathematically analyzed. It is shown that the salinity rise at high latitude as the response to the enhancement of the atmospheric freshwater transport occurs only in thermohaline or haline driven regime. The temperature rise at high latitude as the response to the such change occurs only in thermohaline driven regime or in limited case of haline driven regime. The result of several partial coupled global warming experiments using MRI-CGCM shows the salinity rises and temperature rises in high latitude at the end of the integrations except control run. These results imply that the THC in MRI-CGCM is in the thermohaline driven regime. This consequence well explains the behavior of the THC in glacial periods and in global warming experiment using GCMs.
Appendix

In this section, we give the governing equations of the Rahmstorf’s 4-box model and its equilibrium solutions. By introducing a scale parameter $\gamma$ related to the relaxation times of the system and scaling appropriately, we can obtain the governing equations of the 4-box model for $q > 0$ as follows.

\[
\begin{align*}
\frac{dT_1}{dt} &= q(T_4 - T_1) + \gamma(T_1^* - T_1) \\
\frac{dT_2}{dt} &= q(T_3 - T_2) + \gamma(T_2^* - T_2) \\
\frac{dT_3}{dt} &= q(T_1 - T_3) + \gamma(T_3^* - T_3) \\
\frac{dT_4}{dt} &= q(T_2 - T_4) \\
\frac{dS_1}{dt} &= q(S_4 - S_1) - S_0F_1 \\
\frac{dS_2}{dt} &= q(S_3 - S_2) - S_0F_2 \\
\frac{dS_3}{dt} &= q(S_1 - S_3) - S_0F_3 \\
\frac{dS_4}{dt} &= q(S_2 - S_4) .
\end{align*}
\]

(9)

Similarly, we can write the equations for $q < 0$ as following.

\[
\begin{align*}
\frac{dT_1}{dt} &= -q(T_3 - T_1) + \gamma(T_1^* - T_1) \\
\frac{dT_2}{dt} &= -q(T_4 - T_2) + \gamma(T_2^* - T_2) \\
\frac{dT_3}{dt} &= -q(T_2 - T_3) + \gamma(T_3^* - T_3) \\
\frac{dT_4}{dt} &= -q(T_1 - T_4) \\
\frac{dS_1}{dt} &= -q(S_4 - S_1) - S_0F_1 \\
\frac{dS_2}{dt} &= -q(S_3 - S_2) - S_0F_2 \\
\frac{dS_3}{dt} &= -q(S_1 - S_3) - S_0F_3 \\
\frac{dS_4}{dt} &= -q(S_2 - S_4) .
\end{align*}
\]

(10)

For existence of steady solutions for salinity, a constraint $F_1 + F_2 + F_3 = 0$ is needed. This constraint is automatically satisfied when $F_i$’s are defined by $F_a$ and $F_b$ in Fig. 7.

By putting the terms $d/dt$ into 0 in equations (9) and (10), we can obtain the equilibrium temperature $T_i^*$ and salinity $S_i^*$ for each $i$. Moreover, we can write the equilibrium temperature
difference $\Delta T$ and salinity difference $\Delta S$ between box-2 and box-1 as:

$$
\Delta T = T_2 - T_1 = \begin{cases} 
\left( \frac{T_2^* - T_1^*}{1+3\frac{q^2}{3}+3\frac{q^3}{3^2}} \right) \left( \frac{T_2^* + T_1^* - 2T_1^*}{1-3\frac{q^2}{3}+3\frac{q^3}{3^2}} \right) & \text{for } q > 0 \\
\left( \frac{T_2^* - T_1^*}{1+3\frac{q^2}{3}+3\frac{q^3}{3^2}} \right) \left( \frac{T_2^* + T_1^* - 2T_1^*}{1-3\frac{q^2}{3}+3\frac{q^3}{3^2}} \right) & \text{for } q < 0 
\end{cases}
$$

(11)

and

$$
\Delta S = S_2 - S_1 = \begin{cases} 
\frac{S_2}{q} F_1 & \text{for } q > 0 \\
\frac{S_2}{q} F_2 & \text{for } q < 0 .
\end{cases}
$$

(12)

By substituting (11) and (12) into (1) and putting $\Delta S$ into 0, we can obtain the equation $q + k\alpha \Delta T(q) = 0$. By solving this equation, we can obtain the specific expression for $q_0$ as:

$$
q_0 = -\frac{\gamma}{3} - \frac{\sqrt{2}G}{9\sqrt{H + \sqrt{4G^2 + H^2}}} + \frac{1}{9\sqrt{2}}\sqrt{H + \sqrt{4G^2 + H^2}},
$$

(13)

where

$$
G = 9k\alpha(\gamma (T_2^* - T_1^*) + (T_3^* - T_1^*))
$$

$$
H = (3\gamma)^3 + 9k\alpha(3\gamma)^2T_1^* - 18k\alpha(3\gamma)^2T_2^* + 9k\alpha(3\gamma)^2T_3^* .
$$

References


**Presentations and Publications**

**Original Paper:**


**Conference Report:**


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Radiative Convective Equilibrium Calculations with a Cloud Resolving Model

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Abstract
A new cloud resolving model suitable for climate study is used to calculate radiative-convective equilibrium states that are balanced ones between heat transport due to moist convection and radiative cooling. An appropriate setup for the control experiment is proposed and a number of experiments for the dependency on different physical processes and parameters are performed. In particular, radiative-convective equilibrium states are examined in terms of the bulk coefficients for the surface fluxes and the cloud physics. The dependency on the domain size and resolution is also considered. The radiative-convective equilibrium experiments are necessary for better understandings of the interaction between clouds and radiation that is one of the ambiguous issues of the global warming problem.

Keywords: Radiative convective equilibrium, Cloud resolving model, Cloud-radiation feedback, Cloud physics, Relative humidity

1. Introduction

Radiative-convective equilibrium is not only the basic concept for the understanding of the atmospheric structure, but also the inevitable concept for discussions on climate sensitivity of the global warming problems. In recent years, direct calculations of radiative-convective equilibrium have been performed using cloud resolving models (CRMs) with which the interaction between cumulus convection and radiative process is directly evaluated. The radiative-convective equilibrium calculations with CRMs have the following importance:

1) Estimation of climate sensitivity will be improved by direct evaluation of the cloud and radiation feedback.
2) Properties of physical processes such as cloud physics, radiation, and turbulence schemes used for CRMs can be examined.
3) Cloud parameterization schemes for coarser numerical models can be examined by comparing the results with CRMs.
4) They can be used for evaluation of characteristics of different CRMs.
5) They will lead to the understanding of the tropopause transition layer by directly resolving fine structure near the tropopause.

Nowadays, radiative-convective equilibrium experiments are calculated with CRMs in large domain and long time duration such as about 1000 km x 100 km and several tens of days (Tompkins, 2001). On the other hand, it is reported that the results of radiative-convective equilibrium calculations depend on the choice of the artificial parameters (Tao et al. 1999).

It is preferable to run experiments in larger domain for diagnosis of statistical properties of cumulus convection in radiative-convective equilibrium, since multiple clouds may co-exist within the domain. On one hand, smaller domain experiments have their own values
to save computational cost, because many experiments are required to clarify dependency on external parameters and physical processes. In this study, following Tompkins and Craig (1998), we perform a series of radiative-convective experiments in a relatively smaller domain: 100 km x 100 km. We will discuss statistical properties of equilibrium states and particularly investigate dependencies on surface fluxes and cloud physics. The following calculations are done using a newly developed cloud resolving model (Satoh 2003), which guarantees conservation of mass and energy and is suitable for climate study.

2. Control Experiment

In the first place, we show the result with a control experimental setup as a reference for the following comparisons. The horizontal size is 100 km x 100 km, the top of the domain is 25 km, the horizontal grid interval is \( \Delta x = 2 \) km, the number of vertical layers is 54 with the lowest level at 20 m with stretched layer intervals. We are particularly interested in the impact of cloud physics, so we choose a simple warm rain formula with the bulk method based on the Kessler scheme in this control experiment as a first step toward understanding of behavior of cloud physics. The TKE scheme (Dearendorf) is used as the boundary layer turbulence and the surface fluxes are evaluated with the Louis scheme. We set the minimum speed of the horizontal winds at the lowest level as \( U_{\text{min}} = 4 \) m/s for calculation of surface fluxes. We do not use an interactive radiative scheme for this control experiment, simply, we prescribed radiative cooling rate with a constant value 2 K/day below 9 km and linearly decreasing until zero at 12 km. The initial condition is uniform temperature 250 K at rest, the surface temperature is fixed at 300 K and assumed to be everywhere wet. The integration time is 40 days after 60 days spin-up process.

Figure 1 shows vertical profiles of the domain averaged temperature and relative humidity for the control experiment. This clearly shows the problems of the control experiment: the first is the large temperature gap between the surface temperature and that of the bottom of the atmosphere, and the second is the saturated layer in the upper troposphere between 8-13 km. As a result of the first property, the mean temperature of the atmosphere is very low and the water vapor contained in the atmosphere is very small. The mass weighted mean temperature of this experiment is \( T = 254.9 \) K, and the precipitable water is \( Q = 25.0 \) kg/m\(^2\); these are lower or smaller than the typical values in the tropics: \( T = 260 \) K and \( Q = 60 \) kg/m\(^2\).

Figure 1  Vertical profiles of domain mean temperature (left) and relative humidity (right) for the control experiment at day 100.
We investigated the dependency on the domain size and the grid interval with the same parametric conditions as the control experiment. Three-dimensional calculations with a larger domain 200 km x 200 km using the grid interval $\Delta x = 2$ km and 4 km, and two-dimensional calculations with the domain size of 1000 km and 5000 km are performed. The results indicate that (figures not shown) at the same spacing $\Delta x$ the values of the mean temperature and precipitable water are relatively closer. As $\Delta x$ becomes coarser, the equilibrium states become colder and dryer with larger CAPE.

We investigate statistical properties of the vertical velocity. The probability distribution function (PDF) of the maximum vertical velocity characterizes the equilibrium conditions. As the domain becomes larger, the PDF of the maximum vertical velocity tends to have a peak in the larger vertical velocity side. In other words, this reflects the fact that only a single cumulus convection occurs and multiple cumulus are prohibited in the smaller domain with 100 km x 100 km. This result indicates that the larger domain experiment with 200 km x 200 km is desirable for studying statistical properties of cumulus convection. Nevertheless, since the average values of temperature and precipitable water are not so sensitive to the domain size, we may argue qualitative dependency on the parameters even if the domain 100 km x 100 km is used.

3. Bulk Coefficient

If the total radiative cooling is prescribed, the temperature gap between the surface and the bottom of the atmosphere $\Delta T$ can be estimated using the following energy balance:

$$F_{rad} = Sh + LE$$
$$= \rho C_D V \left[ C_p (T_s - T_0) + L(q_s - q_0) \right]$$
$$\approx \rho C_D V \left[ C_p \Delta T + L(q^*(T_s) - r q^*(T_s - \Delta T)) \right]$$

where $F_{rad}$ is the total radiative cooling rate, $Sh$ is the sensible heat flux from the surface, $E$ is the evaporation flux, and $L$ is the latent heat. It is assumed that the bulk coefficient $C_D$ is the same for the two fluxes. $V$ is the wind speed at the lowest layer of the atmosphere. $T_s$ is the surface temperature, $T_0$ is the temperature at the bottom of the atmosphere, $q_s$ is the specific humidity at the sea surface, $q_0$ is that of the lowest layer of the atmosphere, $r$ is the corresponding relative humidity, and $q^*(T)$ is the saturation specific humidity at temperature $T$. Figure 2 shows how $\Delta T$ varies according to the coefficient $C_D V$ in the case of $r = 80\%$. This shows that in order to obtain a realistic temperature gap $0 < \Delta T < 1$ K, the bulk coefficient must be in the range about 0.01 < $C_D V$ < 0.02, which are relatively larger than accepted values. Figure 3 shows results from radiative-convective equilibrium calculations for different values of the bulk coefficient and the minimum wind speed. The mean temperature and the precipitable water becomes closer to realistic values in the case of larger coefficients $C_D = 0.01$ or $U_{min} = 7$ m/s.

In the case of Tao et al., they introduced a large scale upward motion to obtain a realistic equilibrium state. In order to obtain a state close to the tropical condition seen in the real atmosphere one needs to use either large value of the bulk coefficient, an upward mean velocity, or smaller radiative cooling in such a small domain experiment.
Figure 2  Dependency of the temperature gap $\Delta T$ between the surface and the bottom of the atmosphere on the bulk coefficient $C_D V$ (solid) for $r=80\%$. The dotted curve is for the dry air ($q=0$), and the dashed curve is for the case in which $q_0 = r q_i$ is assumed.

Figure 3  Dependency of the mass weighted mean temperature $T$ and the precipitable water $Q$ on the bulk coefficient $C_D$ and the minimum wind speed $U_{\text{min}}$. CTL denotes the control experiment where the Louis scheme is used with $U_{\text{min}} = 4$ m/s.

4. Cloud Physics

It is found that the saturated layer in the upper troposphere seen in Figure 1 is closely related to the choice of the cloud micro physical scheme. More precisely, it depends on how the conversion rate from the air borne water (cloud) to the sediments (rain/snow) is parameterized. For the control experiment, we use the warm rain scheme in which the autoconversion rate from cloud to rain is represented based on Kessler's formula as

$$AUTO = \tau^{-1} (q_c - q_{cr})$$  \hspace{1cm} (2)

where $\tau$ is the conversion time, $q_c$ is the cloud water content, and $q_{cr}$ is the threshold value of the cloud water: the default values $\tau=1000$ s and $q_{cr} = 1.0$ g/kg are used. If this formula is used, the cloud water floats with air without being converted to rain unless its value exceeds $q_{cr}$. As a result, the cloud water prevails everywhere in the upper layer and it becomes saturated.
The above peculiarity remains the same even if Berry's formula is used for the autoconversion rate from cloud water to rain in the warm rain process. Since $AUTO \propto q^2$, in the case of Berry, thus $\tau \propto q_c^{-1}$; that is, the smaller cloud water stays longer time in the air. Robe and Emanuel (1996) and Grabowski (2003) devised to introduce effects of ice by modifying above conversion time and/or threshold value using the warm rain process. In the case of Robe and Emanuel (1996), $q_{cr}$ linearly decreases below temperature $0^\circ\text{C}$ to zero at $T = -20^\circ\text{C}$. Grabowski (2003) use $q_{cr}=0$ and the faster convection time $\tau = 100\text{s}$. These modifications are thought to be an effect of the ice phase. Grabowski (1998) proposed a simple ice scheme based on three categories of water substance which consist of vapor, air borne condensate, and sediments. Within a certain range of temperature below $0^\circ\text{C}$, the latter two categories are thought as a mixture of liquid and solid phases: cloud water/cloud ice and rain/snow. Besides the conversion to rain is based on Berry's formula, the generation of snow from the cloud ice is calculated using Eq. (2) with $q_{cr} = 0$ and temperature-dependent conversion time: it takes the minimum $\tau = 200\text{s}$ at $-15^\circ\text{C}$.

Figure 4 shows vertical profiles of relative humidity of radiative-convective equilibrium states calculated using the above 4 methods. Although the upper layer is still saturated if Berry's formula is used, this saturation is no longer seen in the case of the other three methods. However, it should be noted that the relative humidity is very different in the middle layer, which results in the difference in the precipitable water $Q$. $Q$ is smallest for the warm rain scheme with Grabowski (2003) modification, and is largest for the simple ice scheme by Grabowski (1998).

Figure 4  Vertical profiles of relative humidity at day 40. Upper left: Berry, upper right: Robe and Emanuel (1996), lower left: Grabowski (2003), lower right: the ice scheme by Grabowski (1998).
5. Summary

Using a newly developed cloud resolving model, radiative-convective equilibrium states are directly calculated in order to study physical processes for the cloud resolving model. Since many kinds of experiments are required to clarify behaviors of the cloud resolving models and of external parameters and physical processes, a relatively smaller domain with 100 km x 100 km is used in three-dimensional calculations. In particular, we examined dependency of the mean temperature and the precipitable water on the bulk coefficient for the surface fluxes and the cloud physical processes. We will further study the dependency on turbulent schemes and more comprehensive ice physical scheme, in order to investigate interaction between clouds and radiation process.

References


Publications and Presentations

Original Papers and Reviews:

Conference Reports:

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Study of the Estimate of New Climate Change Scenarios Based on New Emission Scenarios

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Abstract
In order to perform scenario experiments for the 21st century, the atmospheric concentrations of sulfate aerosols have been calculated for all SRES emission scenarios (A1B, A1T, A1F1, A2, B1, B2) of sulfur dioxide. A chemical transport model MASINGAR developed at the MRI is used for calculating the aerosol concentration. A comparison of the column integrated atmospheric aerosols between those provided by IPCC (2001) and those calculated here reveals some discrepancy in magnitude, suggesting that the atmospheric aerosol burden depends on the chemical transport models to an important extent. With the calculated scenarios, we have started the 21st century scenario experiments. Some preliminary results from the scenario experiments are reported. Surface air temperature rises about 2.4, 2.0, and 2.7 degrees centigrade at the end of this century for A1B, A2, and B2 scenarios, respectively.

Keywords: IPCC, SRES scenario, Radiative forcing, Aerosol, Direct effect

1. Introduction

The purpose of this study is to perform new scenario experiments which are recommended by the Asian-Pacific Integrated Model (AIM) group at the National Institute for Environmental Studies (NIES), the Task Group on Scenarios for Climate Impact Assessment (TGClA) and the Technical Support Unit (TSU) for Intergovernmental Panel on Climate Change-National Greenhouse Gas Inventories Programme (IPCC-NGGIP), for the next IPCC assessment report (AR4). The scenarios are originally provided as emission scenarios and concentration scenarios are additionally provided only for well-mixed greenhouse gases. Under these conditions, the present climate models need atmospheric concentration scenarios for the experiments because a chemical transport model (CTM) is not incorporated in them. Therefore we have calculated the sulfate aerosol distribution by using a CTM developed at the Meteorological Research Institute (MRI), and we are now performing the scenario experiments. This paper will report results from aerosol concentration scenarios calculated with the CTM and preliminary results from new scenario experiments.

2. New Sulfate Distribution Scenarios

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

The MASINGAR includes non-sea-salt (nss) sulfate, carbonaceous, mineral dust and sea salt aerosols but, for the present purpose, accounts only for nss-sulfate aerosols. The sulfur chemistry model treats eight major sulfur compounds, SO$_2$, SO$_4^{2-}$, DMS, H$_2$S, CS$_2$, MSA, DMSO, and DMSO$_2$, and includes seven gas-phase reactions and two aqueous-phase reactions. Calculations of the reactions require the global distributions of oxidants, OH, NO$_3$, H$_2$O$_2$ and O$_3$, which are provided from the observation. The MASINGAR can be operated at several resolutions. In this study, we operate the model at a horizontal resolution of T42, with 30 vertical layers in the hybrid sigma-pressure coordinate, to set the same resolution with the MRI-CGCM 2 (Yukimoto et al, 2002).

2.2 Method of Calculation

The MASINGER can assimilate the dynamical variables with the nudging scheme from JMA global analysis data. In this study, we assimilate temperature and wind velocity from six-hourly three-dimensional atmospheric data calculated by MRI-CGCM 2 with the present radiative forcing, instead of the observed data. Sea surface temperature is given from observed climatology data. We calculate the atmospheric aerosol field for each emission scenario with the industrial sulfur emission data (provided by Dr. Morita). The model also includes natural emissions of SO$_2$ from biomass burning, DMS and CS$_2$ from the ocean, DMS and H$_2$S from terrestrial vegetation and soil, and SO$_2$, SO$_4^{2-}$, and H$_2$S from continuous volcanic activities. We integrate the model for 7 years and make monthly mean from the last 5 years in every 10 years of each scenario.

2.3 Geographical Distribution and Temporal Change in Sulfate Aerosol

Geographical distributions of sulfate burden are calculated with MASINGAR (see Fig. 1). A large seasonal variability can be seen near large emission regions in Fig. 1 for each year although annual mean emissions are used.

Temporal change of global sulfur dioxide emission, sulfate burden, and radiative forcing due to direct effect of sulfate aerosol are shown in Fig. 2. The anthropogenic sulfate burden is 0.79TgS and its direct radiative forcing is -0.54Wm$^{-2}$ in 1990. These values seem to be still large compared with recent studies (IPCC, 2001), and the radiative forcing is as large as that with indirect effect of the first kind (cloud albedo effect) reported in them. In effect the aerosol distribution used in the present scenario experiment may produce a total radiative forcing comparable with those used in the models for IPCC (2001), because the CGCM used for global warming experiment in this study explicitly treats only the direct effect.

3. Global warming experiments with three-dimensional sulfate aerosol distribution

3.1 The Meteorological Research Institute Global Atmosphere-ocean Coupled GCM (MRI-CGCM 2)

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

**Anthropogenic Sulfate Burden**

![Anthropogenic Sulfate Burden](image)

Figure 1  Monthly mean sulfate burden in January (left) and July (right). Top, middle and bottom panels show the distributions in 1990, 2020 and 2050, respectively, for SRES-A1B scenario.
Figure 2 Temporal changes in a) global sulfur dioxide emission for all the six SRES scenarios, A1B, A1FI, A1T, A2, B1 and B2, b) sulfate burden for all the six SRES scenarios, and c) radiative forcing due to direct effect of sulfate aerosol for the SRES A1B, A2 and B2.

3.2 Global Warming Experiments

For greenhouse gases, CO₂, CH₄ and N₂O concentration are provided from SRES tables directly, but other greenhouse gases such as CFCs are substituted by additional CO₂ concentration with an equivalent radiative forcing. The three-dimensional distribution of sulfate aerosol is interpolated in time from the distributions calculated by the CTM. We use the restart data in the year 1990 of the 20th century historical experiments as initial conditions.

Temporal changes in global mean surface air temperature and total precipitation are shown in Fig. 3. Each panel shows the change from the present (from 1961 to 1990). The surface air temperature rises about 2.4, 2.7, and 2.0 degrees centigrade at the end of this century in A1B, A2, and B2 scenarios, respectively. Figure 4 shows the global warming patterns in the year from 2021 to 2050 in A1B (a), and from 2071 to 2100 in A1B (b), A2 (c), and B2 (d). These patterns are qualitatively similar to each other although sulfate aerosol distributions are different among the scenarios. This is because sulfate aerosols do not produce local cooling but rather uniformly reduce the magnitude of global warming anomaly pattern (Noda et al, 2002).
Figure 3 Changes in a) global mean surface air temperature and b) global mean total precipitation.

Surface Air Temperature Change from 1961–1990

Figure 4 Change in surface air temperature relative to the period 1961 to 1990 for a) A1B scenario for the years 2021 to 2050; b) A1B scenario for the years 2071 to 2100; c) A2 scenario for the years 2021 to 2050; d) B2 scenario for the years 2021 to 2050.
4. Summary

In order to perform scenario experiments for the 21st century, the atmospheric concentrations of sulfate aerosols have been calculated for all SRES emission scenarios (A1B, A1T, A1FI, A2, B1, B2) of sulfur dioxide.

A chemical transport model MASINGAR developed at the MRI is used for calculating the aerosol concentration. The MASINGAR includes nss-sulfate, carbonaceous, mineral dust and sea-salt aerosols, although only the nss-sulfate is included for the present calculation, and it accounts for advective transport, subgrid-scale eddy diffusive and convective transport, surface emission, and dry/wet depositions as well as chemical reactions. The advective transport is calculated using a three-dimensional semi-Lagrangian transport scheme. The MASINGAR is coupled with the MRI/JMA98 GCM but, for the present purpose, nudged to atmospheric fields obtained from a simulation with a coupled atmosphere ocean model.

A comparison of the column integrated atmospheric aerosols between those provided by IPCC (2001) and those calculated here reveals some discrepancy in magnitude, suggesting that the atmospheric aerosols depend on the chemical transport models to an important extent.

Since the scenarios are provided in the form of atmospheric concentration, we have started the 21st century scenario experiments. Results from ensemble scenario experiments will be reported in detail in the next report.

References


Presentations

Conference Reports


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2. Atmospheric and Oceanic Environment Modeling
Ultra-high Resolution Modeling of the Tropical Air–Sea Interaction: Natural Variability in Large Domain Cloud Resolving Model

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Abstract
The domain averaged activity of cloud convection in a large cumulus resolving model is found to vacillate with 15–20 day period. By comparing results of experiment performed with the models of different domain size, it is found that the vacillation has larger amplitude and longer time scale in larger domain model. Such vacillation may be understood as a non-linear feedback between convective activity and the energy supply from the ocean.

Keywords: WISHE, Cumulus Convection

1. Introduction
Cumulus convection is one of the most important elements in the energy and hydrological cycle of the earth’s atmosphere. But its smallnesses in temporal and spatial scales compared to the scale of the global atmosphere as a whole prevent it from explicitly calculated in current generation of general circulation model (GCM). Thus every GCM includes cumulus parameterisation, which has many uncertainties that arise from our poor understandings on the interaction between large scale motions and the cumulus clouds. Similar concern also applies to the tropical air–sea interaction (TASI). Usually, the TASI is thought of as a large-scale problem; typical example is El Niño and Southern Oscillation, whose spatial and temporal scales are O(10,000 km) and O(2 years), respectively. Such treatment can partially be justifiable on the grounds that (1) tropical wind system controlling the air-sea exchange processes has large horizontal scale, and that (2) thermal adjustment time of ocean mixed layer as a whole is long. However, atmospheric forcing to the ocean undoubtedly contains considerable smaller scale signals which is produced by cumulus convection. The significance of the small scale components to the TASI problem as a whole is still unclear.

Nakajima (1994) used a two-dimensional cloud model that covers a domain of 16384 km; its high spatial resolution (2 km) and incorporation of cloud micro-physical processes allows cumulus convection to be represented explicitly, and at the same time the domain is as large as the longitudinal size of the Pacific. If we make an ocean numerical model whose resolution and domain size is similar to the cloud model and conduct a coupled experiments, a lot of informations on the TASI in the wide range of resolution will be collected. Nakajima (1998, 1999, 2000) constructed such fine resolution ocean model and examined its ability to simulate time evolution of fresh water layer in the tropical ocean caused by convective rainfall.

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

The focus of the present paper will be placed on the intra-seasonal time scale vacillation reported by Nakajima (2001). The model utilised is basically common to that used by Nakajima
2. Vacillation with Intra-seasonal Time Scale

Figure 1 shows the temporal evolution of the integral properties in the standard run, where the domain size is 32,768 km. All of the quantities vacillate in the time scale of 15 days. The maximum of the kinetic energy occurs around t=15 day, t=30 day, t=45 day, t=65 day. The amplitude of the vacillation relative to the average values are about 30% for evaporation and precipitation, whereas it is as large as 80% for total kinetic energy. The time sequence represents that the vacillation occurs as the following feedback loop: the increase of cloud activity results in strong surface wind (represented by the increase of kinetic energy), which causes the increase of surface flux from the ocean, which causes further increase of convective clouds, after which the static stability in the troposphere increases, and, the convection reduces, after which the reduction of convection causes lowering of the air temperature and recovery of cloud activity.

Figure 1: The temporal evolution of the integral properties in the standard run. From the top to the bottom, (a) average potential temperature anomaly at z=3700 m, (b) average water vapor mixing ratio at z=50 m, (c) total kinetic energy, (d) total evaporation, (e) total precipitation, are plotted. Horizontal axes are time in the unit of hour.
Figure 2 Temporal evolution of rainfall intensity in the standard model. Time goes down, and two cycles of the horizontal domain are shown. Green and red ellipses show enhancement of eastward and westward propagating convective organizations, respectively.

Figure 2 shows the temporal evolution of rainfall intensity. In the domain, there occurs large number of convective clouds. They are organized in a eastward propagating wavy pattern which extends all of the computational domain. The degree of concentration of the clouds varies in time; around the times of maximum kinetic energy, convective activity concentrates into two or three persistent mesoscale systems. Interestingly, one of those mesoscale systems propagates westward, which is in reverse direction that is expected from WISHE theory. Such westward propagation of mesosystem can be attributed to strong westerly surface (disturbance plus basic) wind in those region.

Seven additional model runs are performed starting from slightly different initial conditions. In all of the members (not shown here), similar vacillation of cloud activity occurred. This proves that the 15-20 day scale vacillation of convective activity in the large-domain cloud model is a robust feature.
3. Sensitivity Experiments

Figure 3 compares the amplitudes of the vacillation realized in the runs with different domain sizes. The plotted are normalized total kinetic energy defined as the amount of kinetic energy per 32,768km domain. It is evident that the volume normalized kinetic energy is larger for the larger domain run. Moreover, vacillation becomes more distinct for larger domain; the amplitude relative to the mean value is larger, and the characteristic time scale becomes longer.

Figure 4 compares the amplitudes of the vacillation realized in the runs with different surface wind or cloud microphysics. In general, the vacillation is more prominent in the case with weaker basic state surface wind. It is also noted that kinetic energy reduces when surface wind is very strong. This is because, in that case, the eastward propagating large scale disturbances induced by WISHE is less active.

In case without rain water evaporation, the vacillation is much less active. This should be related to the difference in the mesoscale cloud organization between the runs with and without rain evaporation (see Figure 2 and Figure 5), but the reasons remains unclear at present.

![Total Kinetic Energy](image1)

Figure 3  Temporal evolutions of normalized total kinetic energy in models of different domain sizes. 65,536 km (red), 32,768 km (green), 16,384 km (blue), 8,192 km (purple), 4,096 km (dark olive green), and, 2,048 km (orange).

![Total Kinetic Energy](image2)

Figure 4  Temporal evolutions of normalized total kinetic energy in sensitivity experiments. The standard case(black), the case where evaporation of rain is switched-off(red), the case with the 0.3 m/s basic state wind (blue), the case with the 7.5 m/s basic state wind (green), and, the case with the 12 m/s basic state wind (purple).
4. Concluding Remarks

The enhanced vacillation in large domain runs, especially in the models with “planetary” domain size, suggests that the total amount of convective activity spontaneously vacillate in “intra-seasonal” time scale.

At present, the precise mechanism of the vacillation is unclear. One candidate, suggested from the comparison of the results with different basic state wind, is the nonlinearity in WISHE feedback. That is, when the amplitude of surface wind disturbance exceeds the basic state wind speed, the absolute value of surface wind increases not only to the east of the convective area but also to the west, resulting in the increase of total energy supply from the ocean. Further analysis of the results is necessary to resolve the issue. The sensitivity to the cloud microphysics should also be investigated.
Bantzer and Wallace (1996) shows that various indices in the tropics as a whole, including net precipitation and average temperature, vacillates in the intra-seasonal time scale. The relationship between such vacillation in the real atmosphere and the behavior of idealized two-dimensional model reported here should also be investigated.

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References


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3. Geophysical Fluid Dynamics
Development of Atmospheric General Circulation Model for Terrestrial Planets and Related Fundamental Experiments on the Atmospheric Structures

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Abstract
Aiming for a development of an atmospheric general circulation model (GCM) covering the parameters of the terrestrial planets, we initiate designing an architecture for GCM and performing several implementation tests for the proposed coding scheme, and simultaneously, we are carrying out several basic numerical experiments concerning the atmospheres of Earth and terrestrial planets. As an effort of developing a terrestrial planetary atmospheric model, we are pursuing a desirable software structure of the model which is characterized by high readability of its source code and high flexibility of changing the experimental design and the implemented physical processes. Implementation tests are performed with a spherical shallow water spectral model as a pilot system of the dynamical core of our GCM. As a basic experiment concerning the terrestrial planetary atmospheres, a parameter study on the value of solar constant is performed with the aqua-planet situation on the condition that surface albedo changes according to surface temperature. The results provide a GCM version of the climate regime diagram which is formerly discussed with an energy balance model. As a basic experiment concerning the Earth-like conditions, a parameter study is performed considering the effect of two areas of warm sea surface temperature (SST) placed at the equator of an aqua-planet with a zonally uniform and equatorially symmetric SST distribution. It is suggested that the precipitation anomaly pattern can be described by the equatorial Kelvin-wave/Rossby-wave structure associated with moisture convergence/divergence.

Keywords: GCM, Planetary atmosphere, Code readability, Aqua planet, Runaway greenhouse state, Large ice cap instability, Small ice cap instability, Warm SST anomaly, Equatorial precipitation, Tropical circulation

1. Background

For the purpose of recognizing the climatic state of Earth and its variability, we are trying to introduce a stance of comparative planetary science ranging over Mars, Venus, and Earth, and consider the terrestrial climate from this viewpoint. Understanding of the climates of Mars and Venus have been developed by applying the accumulation of knowledge on the terrestrial climate. Knowledge on the terrestrial climate can be tested and improved through its application to the planetary atmospheres. On the other hand, consideration of the Martian and Venusean climates provides us new viewpoints for recognizing the terrestrial climate. The cold climate of
present Mars reminds us of the recently proposed hypothesis that Earth might have experienced a cold climate and been covered with ice globally in neo-Proterozoic (Snowball Earth hypothesis; Hoffman et al., 1998). It is also reported that the ancient Mars was in a warm climate state and had an ocean on its surface (Carr, 1996). The issue of the appearance of cold climate or warm climate is the common problem for both of the terrestrial and Martian climates. The thick atmosphere and high surface temperature of present Venus is considered to be caused by the occurrence of the runaway greenhouse state in the past. Present terrestrial atmosphere, if only the tropics are considered, is under the condition that the runaway greenhouse state can appear. Consideration of the condition for the appearance of the runaway greenhouse state contributes not only to the investigation of the evolution of Venusian atmosphere but also to the understanding of the present terrestrial climate and its stability.

In order to ensure the stance of comparative planetary science, it is necessary to acquire a model which can simulate various climatic states of terrestrial planetary atmospheres; the model enables us to consider the present climatic states of Earth, Venus, and Mars, and the runaway greenhouse state and the ice-covered state. The substance of our activity is to make efforts toward construction of such a model and to continue considerations on Earth and other terrestrial planetary atmospheres by performing basic numerical experiments in the similar way as conducted so far by ourselves. Those basic numerical experiments contribute not only to revealing dynamical structures embedded in the climates of Earth and other planets, but to providing guidelines for constructing the terrestrial planetary atmospheric model. In order to construct a model that can calculate a climatic state for which poor or no observational data is obtained, it is essential to accumulate various kinds of clues by which the model validity can be examined.

For those purposes described above, we are now engaged in the following three major activities:

- **Implementation tests toward the development of GCM for terrestrial planetary atmospheres:**
  Desirable software structure of the model and appropriate data structure for the model output are being considered. In the considerations, experiences obtained through the following basic experiments concerning the terrestrial atmospheres are incorporated. Implementation tests of those structures are performed. Those results obtained in 2002 are reported in section 2.

- **Basic experiments concerning terrestrial planetary atmospheres:**
  Multiple solutions of the climate system, such as ice-covered state, partially ice-covered state, the runaway greenhouse state, are being considered by performing a parameter study on the value of solar constant. These experiments are also aimed to acquire computational technologies for calculating extreme climate conditions. The results obtained in 2002 are reported in section 3.

- **Basic experiments under Earth like condition:**
  The atmospheric responses to the existence of warm sea surface temperature areas are investigated by performing experiments under the aqua planet condition. Issues on the tropical precipitation structures such as Madden-Julian oscillation, super cloud cluster, and cloud cluster are being considered. 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. The results obtained in 2002 are reported in section 4.
2. Implementation Tests Toward the Development of GCM for Terrestrial Planetary Atmospheres

In constructing GCM for planetary atmospheres, we focus on the following three points on the structure of software:

1. **Flexibility of changing the schemes implemented in the model**
   In order to cover the parameter range of the terrestrial planets, model should be flexible in changing the experimental design and the implemented physical processes. We are pursuing a desirable software structure of the model with high flexibility.

2. **Readability of the source code**
   In order to describe and understand the results obtained under various conditions, source code of the model should be readable like as equations written in the textbooks or papers. We are pursuing a desirable coding style with high readability.

3. **Data portability**
   In order to handle the data at remote supercomputers efficiently, the data produced by the model should be transparent through network. Moreover, data which is a number of huge three dimensional time series should have suitable structure for analysis and visualization.

As for the data structure, we have adopted netCDF (network Common Data Form) file format so as to ensure the transparency through network, and are developing gtool4 netCDF convention (GFD-Dennou Club, 2000) which embodies a data structure for analysis and visualization. As for the coding design, we are developing Hierarchical GFD Spectral Models (SPMODEL; GFD-Dennou Club, 2001). In SPMODEL, with supplying functions for spectrum transformation and spatial differentiation, information hiding is realized. This software structure enlarges readability of its source code and flexibility of changing the experimental design and the physical processes.

We are now developing a spherical shallow water spectrum model as a pilot system of the dynamical core of our GCM. Time integration is performed by the leapfrog scheme. To remove the computational modes, time filter of Asselin (1972) is applied. The expression of time integration of vorticity in source code is as follows:

```fortran
  do it=1, nt
    w_ZetaA = w_ZetaB + 2 * delt * &
            (- w_Div_xy_xy((xy_Col1 + xy_w(w_Zeta)) &
              * xy_w(w_U), &
            (xy_Col1 + xy_w(w_Zeta)) &
            * xy_w(w_V)) / R0)
  end do
```

where `w_Div_xy_xy( )`, `xy_w( )` are functions for two-dimensional divergence and spectrum inverse transformation, respectively. Adopting SPMODEL enables us to correspond source code to original model equations easily and makes source code readable.

Figure 1 shows the result of performance test: shallow water model test case 5 suggested by Williamson et al. (1992). The increase of computational cost brought by massive usage of array functions is negligible small.
Figure 1  The performance test of spherical shallow water model.  Figures show the response of relative vorticity to a mountain at (a) initial state, (b) 5 day, (c) 10 day, (d) 15 day, respectively. Resolution of the model is T106.

3. Climate Dependency on Solar Constant: Snowball Earth

As a basic experiment concerning the terrestrial planetary atmospheres, a parameter study on the value of solar constant is performed. The aim of this study is to examine the occurrence conditions for ice-covered state, runaway greenhouse state, and to investigate the atmospheric structure that causes such extreme climate states. The dependency of climate on solar constant is formerly discussed with an energy balance model (Budyko, 1969; Sellers, 1969). We reconsider the multiple equilibrium solutions, the large ice cap instability, and the small ice cap instability with the use of a GCM which expresses the atmospheric circulation explicitly and can describe the runaway greenhouse state. The occurrence of the large ice cap instability in GCM is reported last year. This year, we discuss the small ice cap instability and its correspondence in GCM.

The model used is AGCM5 of GFD Dennou Club edition, which is the three-dimensional primitive system on a sphere (Swamp Project, 1998). The model is basically the same as that developed by and used in Numaguti (1993). The dynamical part is represented by the pseudo spectral method with the triangular truncation at wavenumber 21 (T21) and by the sigma vertical coordinate with 16 or 32 vertical levels. The model contains the following simplified physical processes. The atmosphere consists of a condensible component (water vapor) and a noncondensible component (dry air). Only water vapor absorbs and emits longwave radiation. The absorption coefficient of water vapor is constant and independent of wavelength. Dry air is assumed to be transparent. The radiative effects of clouds and the scattering of radiation are excluded. The employed cumulus parameterization is the adjustment scheme. The vertical diffusion is represented by Yamada-Meller Level II scheme. All of the surface is assumed to be swamp ocean; heat capacity is zero and wetness is unity. Heat transport by the ocean is
excluded. The surface albedo of the region whose surface temperature is below the freezing point, 263K, is 0.5. Otherwise, the surface albedo is zero. The distribution of incoming solar flux is given by the annual and daily average of that evaluated with the present terrestrial orbital parameters. The value of solar constant $S$ is varied in a range between 1200 W/m$^2$ and 2000 W/m$^2$.

The relationship between solar constant and ice line latitude obtained by our GCM is shown in Figure 2. In this figure, all of the results starting from different initial conditions are plotted. The partially ice-covered statistically equilibrium states obtained under $1300 \leq S \leq 1570$ W/m$^2$ are from the initial condition of the isothermal atmosphere (280K), and those obtained under $1250 \leq S < 1300$ W/m$^2$ are given by decreasing the value of solar constant gradually starting from partially ice-covered state calculated with $S = 1300$ W/m$^2$. As reported in 2002, the ice line latitude of the partially ice-covered state with the most spreading ice area obtained by the GCM is located at 22 degree. No partially ice-covered state with ice line latitude lower than 22 degree is obtained. The results suggest that the large ice cap instability occurs also in the three-dimensional system as in the energy balance model.

![Figure 2](image)

**Figure 2**  Relationship between solar constant and ice line latitude obtained by the GCM. ○, ●, and × represent an ice-covered equilibrium state, a partially ice-covered or an ice-free equilibrium state, and a runaway greenhouse state, respectively. Suffix “F” and “R” represent the results whose initial states are an ice-covered state calculated with $S = 1000$ W/m$^2$, and a runaway greenhouse state obtained with $S = 1600$ W/m$^2$, respectively. Suffix “P” represents the results obtained with decreasing $S$ gradually starting from a partially ice-covered state obtained with $S = 1300$ W/m$^2$. Marks with no suffix represent the results whose initial state is the isothermal state.

As for the small ice cap side, our GCM yields statistically equilibrium states. This might suggest that the small ice cap instability does not exist in our three-dimensional system. However, in the GCM results, the ice line latitude does not stand still at a fixed latitude but oscillates. The amplitude of the latitudinal variation of the ice line increases with the increase of $S$ (Figure
3). This might suggest that the small ice cap instability exists and around the unstable solution there is a stable periodic orbit; the GCM correspondence of the occurrence of the small ice cap instability of the energy balance model might be a Hopf bifurcation.

![Image of ice edge latitudes](image)

**Figure 3** Time series of the ice line latitudes obtained by the GCM. Plotted are the averaged latitudes of the northern and southern lines. The upper, middle, and lower lines correspond to $S = 1560, 1380, 1250$ W/m$^2$, respectively.

4. **Aqua-planet Warm SST Experiment: Two Warm SST Areas at the Equator**

As a basic experiment under Earth like condition, we have been investigated the response of global precipitation distribution and circulation structure to warm SST anomaly area by the use of GCM. Last year, we showed that precipitation increase to the east of warm SST area is caused by moisture convergence associated with Kelvin wave-like response and drying to the west of warm SST area is caused by the moisture divergence at the equator associated with the Rossby pressure response for the case with a warm SST anomaly area placed at the equator of zonally uniform and equatorial symmetric SST distribution (Hosaka, et al., 1998; Toyoda et al., 2003). This year, the responses to the existence of two warm SST area placed at the equator are investigated. In this case, it is unclear that convection centers over warm SST areas have the same intensity. There is a possibility that, unless the two warm areas are placed symmetrically with the distance of 180°, precipitation at the eastern area is significantly intensified compared to that over the western area, because of the moist convergence produced by the precipitation peak on the western warm area. There is also a possibility that, with the same reason, oscillation of the precipitation intensity occurs between the two warm areas. In the followings, we show the intensity of convection centers, the east-west asymmetry around the warm SST areas, and behavior of precipitation structures.

The model used is AGCM5.3 of GFD Dennou Club edition, and basically same as the model used in section 3 except for the radiation scheme and surface condition. The employed
cumulus parameterization is the adjustment scheme. The absorption and scattering of the solar radiation are not included. The longwave absorption is represented by four spectral bands, three of which are of water vapor and one is of dry air. The scattering of the longwave radiation is not included. The absorption coefficients of the longwave radiation are chosen so that the cooling profile of the atmosphere roughly resembles the observed one. Spatial resolution is T42L16. The SST distribution consists of two patches of localized warm SST anomaly superimposed on the equatorially symmetric, zonally uniform basic SST distribution. A local warm SST area is rectangular shape with a longitudinal extent of 40° and a latitudinal extent of 20°. The peak value of the SST anomaly is 4K; the actual maximum value achieved at the grid point is 3.4K. Centers of the equatorial local warm SST areas are placed at the longitude of 180° and at the longitude of 215°. The initial condition of the experiment is the state at 700 day of standard experiment without warm SST anomaly area. Integration is performed for 1000 day.

Figure 4 shows the distribution of precipitation anomaly obtained in the experiment. The atmospheric response around the warm SST area can be understood with the discussions of Toyoda et al. (2003); There appear equatorial Kelvin-wave like structure associated with moisture convergence to the east of the warm SST area, and Rossby-wave like structure associated with moisture divergence to the west of warm SST area. The increase of precipitation to the east of warm SST area is accompanied with negative surface pressure anomaly and positive mid-tropospheric temperature anomaly, which confirm the emergence of Kelvin-wave like response.

![Figure 4](image_url)  
**Figure 4  Precipitation distribution of the case with two warm SST anomaly areas.** Difference from the result of standard experiment without warm SST area is shown. The data of last 700 day are used. Unit is W/m².

Figure 5 shows the time evolution of equatorial precipitation. As for the intensity of precipitation at convection centers, the intensification of precipitation over the western warm SST area exists but weak. There does not appear a clear signal of oscillation of the convection center intensity. In Figure 5, we recognize grid-scale eastward propagating precipitation structures over the warm SST areas, while westward propagating grid-size precipitation structures over the other area. Eastward moving grid-scale precipitation structures can be recognized as super cloud cluster. The eastward moving structures are accompanied with negative surface pressure anomaly and positive surface zonal wind (figures not shown). The reason why the grid-size precipitation is organized to move eastward only at the warm SST areas has not been clarified
yet.

![Figure 5](image-url)  
**Figure 5  Time evolution of equatorial precipitation.** The difference from the result of standard experiment without warm SST area is shown. Unit is W/m².

**References**


http://dennou-t.ms.u-tokyo.ac.jp/arch/gtool4/gt4ncconv-current/SIGEN.htm

http://dennou-t.gfddennou.org/arch/spmodel/index.htm


Publication


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Numerical Simulations for Cyclone Generated Around a Large Mountain

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Abstract
A great number of simulations that have a lot of Fr, Ro, and aspect ratio of the mountain, were executed in order to focus the generation of cyclonic vortex in upper and lower layer near mountain in a rotating stratified flow system with the 3-D nonhydrostatic meteorological prognostic model. In the upper layer, the generation of cyclonic vortex is characterized mainly by an air column stretching, and also seems to be concerned with Fr, Re, the scale of the mountain, and the height of the mountain height. A quasigeostrophic equation with the idea of the top layer theory leads to be inversely proportional to square of Fr and our result is roughly in accordance with this theory. In the lower (middle) layer, generation of vortex seems to be concerned with the condition of baroclinicity in the leeward region of the mountain. With the consideration of the baroclinic instability, if the condition in the leeward region of the mountain is the baroclinic instability, the baroclinicity dominates and the wave that has three-dimensional structure is dominant, on the other hand if the condition is not the baroclinic instability, lee vortex in the middle layer has standing structure.

Keywords: Rotating stratified system, Numerical model, Wake, Lee vortex, Lee wave, Karman Vortex, Vortex generation

1. Introduction

To investigate the mountain’s effects on the vortex generation behind a great mountain, it is necessary to understand the global scale fluid dynamical phenomena with rotating, stratification, and high Reynolds number effects. There are a great number of theoretical, analytical, and numerical studies on the interaction between the mountain and stratified flow, and these studies concern with the idea of Froude number for high Re number. Some of stratified effects are lee wave and lee vortex. Hunt et al. (1997 and 2001) suggested that the regime of the flow that surrounds the mountain can be separated to two flow regimes. One of them is the flow “over the mountain” that concerns with lee waves, and another of them is the flow “around the mountain” that concerns with lee vortices. These flows form the two layers, “top layer” and “middle layer”. Top layer is the region that contains the crest of the mountain and is dominated by lee waves, and middle layer is below the top layer and is dominated by lee vortices or wake. So, within the area surrounding the mountain, some kinds of vortices exist with inclined axes corresponding lee vortex and lee wave.

On the other hand, with rotational background, studies on the formation of initial vortex are not adequately. In this system, some kinds of lee wave come to contain vortex with vertical axis because of the conservation of quasi-geostrophic potential vorticity. Huppert and Bryan (1976) showed the theoretical and numerical discussion for the initial formation of vortices due to the stretching of air column over the crest with free slip boundary condition. This initial generating vortex cannot be considered with non-rotating system because of the lack of “Coriolis vorticity”. However, the theory of them seems not to have the effect of the
lee wave concepts because they executed a lot of numerical experiment with coarse domain, so lee wave effects were difficult to confirm. It is reasonable to connect the theory of Huppert and Bryan (1976) and of Hunt et al. (2001) for more detail analysis for the initial generating vortex and its characteristics.

We examined what parameters decide the strength of the initial generating vortex and the interaction lee wave to lee vortex on the rotating stratified system with 3-D non-hydrostatic meteorological prognostic model (Sha et al., 1996 and 1998). In section 2, we introduce the summary of our numerical model and of numerical experiment. The initial generated vortex in the top layer is discussed in section 3, the discussion of the lee vortex in the middle layer is contained in section 4, and in section 5, we show the summary of this paper.

2. Model Description and Outline of Our Numerical Experiments

The initial generated vortex and lee vortex were performed in the three-dimensional non-hydrostatic meteorological prognostic model (Sha et al., 1996 and 1998), which contains the anelastic, primitive equations for predicting with time integral these equations. The shape of the mountain we introduced is represented in Table 1. In Table 1, hm means the height of the crest, L represents the horizontal mountain scale, and \((x_0, y_0)\) depicts the center of the mountain. Coordinate system is non-equally gridded (100 m-2000 m) terrain following for vertical, equally gridded \((x_{50} \text{ km} \ y_{50} \text{ km})\) for horizontal, and total domain is \(x_{6000} \text{ km} \ y_{6000} \text{ km} \ z_{25} \text{ km}\) with dumping layer for the height upper than \(15 \text{ km}\), so total grid is \(x_{120} \ y_{120} \ z_{44}\). For detail description for dynamical pressure, 3-D Poisson equation is adopted with the Neumann boundary condition (Schumann and Volkert, 1984). Upper boundary condition is fixed to the initial value for wind and temperature, and lower boundary condition is fixed to no-slip and also the initial circumstance for each parameters. Vertical eddy diffusive coefficient of momentum is solved by Blackadar (1962), and as for that of heat, Plandtl number is used. Values of the windward boundary are also fixed to initial ones, on the other hand, other three lateral boundary conditions are radiation condition. The initial values of wind and thermal stability are identical value and Coriolis forcing is fixed to \(f_0\)-plane value, and numerical integration is ranged from “impulsive start”, the time initial condition, to \(Ut/L \sim 25\) as non-dimensional time scale. Run's parameters are wind speed \(U\), Brunt-Väisälä frequency for stability, \(N\), Coriolis forcing \(f\), mountain height \(hm\), and horizontal scale \(L\). We executed two type simulations, one is that \(U\) and \(N\) are variable parameter each other and is fixed that \(f = 8.338 \times 10^{-5}\ 1/s\), \(hm = 4000\ m\), and \(L = 500\ km\). Another is that \(f\), \(hm\), and \(L\) are variable and other parameters are fixed as \(U = 10\ m/s\), \(N = 0.02\ 1/s\). These are summarized to Table 2 and Table 3.

Table 1 Mountain shape.

| \(h = h_s \exp\left\{-1.0 \times \frac{(x-x_s)^2 + (y-y_s)^2}{L^2}\right\}\) |

Table 2 Sweep parameters for \(U\) and \(N\).

<table>
<thead>
<tr>
<th>(U\ (m/s))</th>
<th>3</th>
<th>5</th>
<th>6</th>
<th>8</th>
<th>10</th>
<th>12</th>
<th>14</th>
<th>15</th>
<th>16</th>
<th>18</th>
<th>20</th>
<th>22</th>
<th>24</th>
<th>26</th>
<th>28</th>
<th>30</th>
<th>32</th>
<th>36</th>
<th>40</th>
</tr>
</thead>
<tbody>
<tr>
<td>(N\ (1/s))</td>
<td>0.005</td>
<td>0.010</td>
<td>0.015</td>
<td>0.020</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>
### Table 3  Sweep parameters for Coriolis, height and horizontal scale of the mountain.

<table>
<thead>
<tr>
<th>Coriolis parameter (f)</th>
<th>hm (m)</th>
<th>L (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>2000</td>
<td>100</td>
</tr>
<tr>
<td>0.05f₀</td>
<td>3000</td>
<td>200</td>
</tr>
<tr>
<td>0.1f₀</td>
<td>5000</td>
<td>250</td>
</tr>
<tr>
<td>0.2f₀</td>
<td>6000</td>
<td>400</td>
</tr>
<tr>
<td>0.4f₀</td>
<td>7000</td>
<td>750</td>
</tr>
<tr>
<td>0.6f₀</td>
<td>8000</td>
<td></td>
</tr>
<tr>
<td>0.8f₀</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.2f₀</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.4f₀</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.6f₀</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

### 3. Initial Generating Vortex in Top Layer

First, the result of $U = 10$ m/s and $N = 0.01$ 1/s run (RUN 1) are shown in Figure 1(a)-(c). Figure 1(a) depicts wind vectors and of vorticity of vertical component contours at the non-dimensional time 0.6 after the impulsive start, and windward is left side of this figure. When the flow from windward goes over (or around) the mountain, the streamline is bended to left with looking leeward. This means that the air column moving from windward that goes over the crest is suppressed vertically because of the air column comes to be caught between the mountain’s shape and the top of the atmosphere, so with the conservation of potential vorticity the air column comes to have an anticyclonic vorticity. On the other hand, when this column goes down the slope of the mountain, it comes to stretch and to take a cyclonic vorticity, so the streamline of the flow is bended to right with looking leeward. Vertical winds and potential temperatures are shown in Figure 2 (a). In Figure 2 (a), there are the colder region on the windward of the mountain and the warmer region on the leeward of the mountain. This means that the pressure is higher at colder region and the flow bends to left with looking downward because of geostrophic balance, whereas the pressure at warmer

![Figure 1](image.png)

**Figure 1**  RUN 1 wind and vorticity on the height of 4000 m at $\frac{tU}{L} =$ (a)0.6, (b)1.8, and (c)3.0.
Figure 2  RUN 1 on the height of 4500 m at tU/L = 0.6 results. (a)potential temperature and vertical wind (thin line means upwind, while thin broken line means downwind by every 1.0 cm/s, and thick line means potential temperature by every 1.0 K), (b)time gradient term of vertical vorticity equation (line means positive value and broken line means negative value, with every 0.5*10^{-3} 1/s^2), (c)horizontal advection term of vertical vorticity equation (every 1.0*10^{-3} 1/s^2), (d)divergence term of vertical vorticity equation (every 1.0*10^{-3} 1/s^2), (e)tilting term of vertical vorticity equation (every 0.5*10^{-3} 1/s^2), and (f)friction term of vertical vorticity equation (every 1.0*10^{-3} 1/s^2).

Table 4  Vorticity equations for three components.

<table>
<thead>
<tr>
<th>$\xi = \frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}$</th>
<th>$\eta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}$</th>
<th>$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\frac{\partial \xi}{\partial t} = -u \frac{\partial \xi}{\partial x} - v \frac{\partial \xi}{\partial y} + \frac{\partial \xi}{\partial z} - \xi \left( \frac{\partial w}{\partial y} + \frac{\partial v}{\partial z} \right) + \frac{\partial v}{\partial x} \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \frac{\partial u}{\partial y} + \frac{\partial w}{\partial z} \frac{\partial u}{\partial x} + f \frac{\partial w}{\partial z} + \frac{1}{\rho \theta} \left( \frac{\partial \theta}{\partial y} \frac{\partial \xi}{\partial z} - \frac{\partial \theta}{\partial z} \frac{\partial \xi}{\partial y} \right) - F_x$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\frac{\partial \eta}{\partial t} = -u \frac{\partial \eta}{\partial x} - v \frac{\partial \eta}{\partial y} + \frac{\partial \eta}{\partial z} - \eta \left( \frac{\partial w}{\partial x} + \frac{\partial u}{\partial z} \right) + \frac{\partial v}{\partial y} \frac{\partial u}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} + \frac{\partial w}{\partial x} \frac{\partial u}{\partial z} + f \frac{\partial w}{\partial x} + \frac{1}{\rho \theta} \left( \frac{\partial \theta}{\partial x} \frac{\partial \eta}{\partial z} - \frac{\partial \theta}{\partial z} \frac{\partial \eta}{\partial x} \right) - F_y$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\frac{\partial \zeta}{\partial t} = -u \frac{\partial \zeta}{\partial x} - v \frac{\partial \zeta}{\partial y} + \frac{\partial \zeta}{\partial z} - \zeta \left( \frac{\partial w}{\partial x} + \frac{\partial v}{\partial y} \right) + \frac{\partial v}{\partial z} \frac{\partial w}{\partial x} - \frac{\partial w}{\partial z} \frac{\partial v}{\partial x} + \frac{\partial w}{\partial y} \frac{\partial v}{\partial x} + f \frac{\partial w}{\partial y} + \frac{1}{\rho \theta} \left( \frac{\partial \theta}{\partial x} \frac{\partial \zeta}{\partial y} - \frac{\partial \theta}{\partial y} \frac{\partial \zeta}{\partial x} \right) - F_z$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

region is lower and the flow bends right. Figure 2 (b)-(f) shows each terms of the vorticity equation for vertical component at same time and same height of Figure 1. Table 4 represents the vorticity equation we used. Except for advection term, divergence term is most dominant, and the second dominant term is friction term that works inversely effective. Tilting term means to convert vertical vorticity to horizontal vorticity. So, the initial vortex generation is
formed by stretching and simultaneously weakened by friction and tilting effects. Other runs are shown in Figure 3. RUN 1, RUN 2, and RUN 3 in Figure 3 mean three runs, RUN 1 is already described, RUN 2 means $U = 15 \text{ m/s}$, $N = 0.01 \text { 1/s}$, and RUN 3 represents $U = 20 \text{ m/s}$, $N = 0.01 \text { 1/s}$ results at $tU/L = 3.0$. From these runs stretching effect is the main source of the initial generating vortex. However, these vorticities are not same values each other but seem to depend on the wind speed and the stability. Maximum value of positive vorticity behind the mountain on RUN1 is $65*10^{-6} \text{ 1/s}$, one on RUN2 is $83*10^{-6} \text{ 1/s}$, and one on RUN 3 is $79*10^{-6} \text{ 1/s}$. With considering the conservation of the potential vorticity, the vorticity of the initial generating vortex seems to depend on the stability and Coriolis parameter, and also the height of the mountain as the strength of the stretching. So on the next step, we examine these correlations between the value of vorticity of the initial generation and several parameters that are fixed from each runs described in Table 2 and Table 3. Figure 4 shows this relationship with $Fr = U/Ohm$ parameter, and the relationship between the non-dimensional vorticity and RUN’s

![Figure 3](image_url)

**Figure 3** Wind and vorticity on the height of 4000 m at $tU/L = 3.0$ of the RUNs of (a)$U = 10\text{ m/s}$, $N = 0.01 \text { 1/s}$, (b)$U = 15 \text{ m/s}$, $N = 0.01 \text { 1/s}$, and (c)$U = 20 \text{ m/s}$, $N = 0.01 \text { 1/s}$.

![Figure 4](image_url)

**Figure 4** Non-dimensional vorticity of various RUNs with Fr.
parameters decided subjectively is represented in Table 5. If Fr is less than 0.3, non-dimensional vorticity is proportional to square of Fr and some other parameters, and if Fr is larger than 0.3, non-dimensional vorticity is inversely proportional to square of Fr and other parameters. One of the most important results is that this relationship does not contain Ro or Coriolis parameter. We estimated this relationship with Huppert and Bryan (1976)’s quasi-geostrophic theorem and with the idea that Hunt et al. (2001)’s “top layer” and “middle layer” theorem. The result of the theoretical relationship is shown in Table 6 and this relationship by quasi-geostrophic theorem meets some corresponding to the relationship between non-dimensional vorticity and Fr within 0.3 < Fr < 0.9, but more work is necessary to have better corresponding between the numerical result and quasi-geostrophic theorem for aspect ratio and Re, etc.

**Table 5 Non-dimensional vorticity relationship.**

\[
\frac{\zeta}{U} = \zeta^* \approx Fr^2 \text{Re}^{-\frac{1}{2}} \left( \frac{hm}{L} \right)^{-1} \left( \frac{hm}{H} \right) \text{ for } Fr \leq 0.3,
\]

\[
\frac{\zeta}{U} = \zeta^* \approx Fr^2 \text{Re}^{-\frac{1}{2}} \left( \frac{hm}{L} \right)^{-1} \left( \frac{hm}{H} \right) \text{ for } 0.3 \leq Fr \leq 0.9.
\]

**Table 6 Non-dimensional vorticity relationship from**

\[
\frac{\zeta_5,7L}{U} = \frac{1}{2} Fr + (1 - Fr) \log(1 - Fr) \approx \frac{1}{2} Fr^{-\frac{1}{2}}.
\]

### 4. Vortex Generation in the Middle Layer

In this section, we discuss how the vortex maintains in the middle layer by considering dynamical mechanism of this. In stratified flow, the formation of the wake or Karman vortex depends on Re and Fr. Our experiments are under high Re and Coriolis effect, so these parameters seem to have some of the effects on phenomena in the middle layer. Especially Ro effect makes the difference of the flow speed between around left hand side and around right hand side of the mountain, so these horizontal shears may generate some kinds of the wake deformation. And due to geostrophic balance, there is horizontal gradient of temperature by vertical wind shear, and this may change some kinds of dynamical instability in the region behind the mountain. So, the characteristics of dynamical stability are discussed by the vortex formation of several runs.

Figure 5 shows the flow and vertical vorticity pattern of RUN 4 (U = 18 m/s, N = 0.015 1/s) on the height of 2000 m, this height is included in the middle layer, at tU/L = 4.9, 6.7, and 8.6. After the initial generating cyclone go leeward far from the mountain, other cyclonic and anticyclonic vorticities are generated near the lateral of the mountain and move to leeward.
the same Figure 5 but for RUN 5 (U = 6 m/s, N = 0.005 1/s), does not express any moving cyclone-anticyclone pairs because of its stationary flow pattern nevertheless these two runs have the same value of Fr = 0.3. This means that it is difficult for vortex to generate and to maintain behind of the mountain.

Figure 5  RUN 4 wind and vorticity on the height of 2000 m at tU/L = (a)4.9, (b)6.7, and (c)8.6.

Figure 6  RUN 5 wind and vorticity on the height of 2000 m at tU/L = (a)4.8, (b)6.6, and (c)8.4.

So, to clarify the vortex generation and its dependence of the flow, Coriolis, and mountain parameters, we defined the existence of the vortex as Table 7. i and j means the unit vectors that are streamwise in x and spanwise in y, respectively. If any D_k > 0 at a point, cyclonic vortex is exist at that point, and if any D_k < 0 at a point, it means anticyclonic vortex is formed at that point. We adopted value of n as 8 after the time of tU/L > 3.6 in all of our runs. With this method, many vortices that have various lifetimes are appeared, so we classified these vortices by those lifetimes, and these results are shown in Figure 7. One of any kinds of the character means one of our runs as representing Fr and Ro, and some runs are overwritten because of having the same Ro and Fr but difference runs each other. The character "o" means the existence of the vortex that had a long life, the character "d" represents the vortex existence having a short life, and "x" means no vortex existence. By Figure 7, a low Fr run has a large possibility of the vortex formation but if Ro is not small value, the possibility become less, so Coriolis effect seems to work to suppress the formation of the vortex in the middle layer. In a low Fr situation, the top layer is thin and lee wave is not dominant, but the middle layer is thick, so the width of the lee vortex seems to be thick. On the other hand, if Fr
is not a low value, lee wave in the top layer becomes dominant and lee vortex comes to be slightly. With considering that lee wave as the vortex has the horizontal axis and lee vortex have the vertical axis, and lee wave contains Coriolis effect as inertia gravity wave having a tilted axis, if Fr is low, the interaction between lee wave and lee vortex is little, so lee vortex in the middle layer still can have vertical axis of vortex. On the other hand, if Fr is not low, the effect of lee wave is greater in the leeward region of the mountain, so the formation of the lee vortex seems to be difficult.

Table 7 Definition of vortex formation.

\[ u(x, y) = 0, \]
\[ \text{and} \]
\[ D_k(x, y) = \frac{\delta(k \times u(x, y)) \cdot r_k}{\delta \cdot r_k} \neq 0, \]
\[ r_k = i \cos \theta_k + j \sin \theta_k, \]
\[ \theta_k = \frac{2\pi k}{n} \quad (k = 1, 2, \cdots, n - 1). \]

Figure 7 Diagram of vortex existence for each run. "O" means vortex have a longer lifetime than t/U/L = 4.0, and move more than L before diffusion. "a" means vortex have a longer lifetime than t/U/L = 4.0 but is not enough to move than L. "d" represents vortex have a little longer lifetime than t/U/L = 1.0. "x" denotes no vortex existed or existing vortex have no long lifetime for t/U/L = 1.0. Thick line means the critical line of baroclinic instability. right region of this means baroclinic instability with the scale we decided (for more detail, see the text).
So we introduced the baroclinic instability consideration, that originates from stratified and baroclinicity, as the vortex formation of dynamical characteristics in the leeward of the mountain. Thick line in Figure 7 means the baroclinic instability curve with some approximations in leeward of the mountain, constant vertical wind shear, thickness of baroclinic layer is equal to that of the middle layer, and horizontal scale of approximated phenomenon is equal to the scale of cross section of the mountain. Right region of this line means baroclinic instability within leeward region of the mountain at that run, on the other hand left region of this line means that it is not unstable for the perturbation that has the tilted axis of the vortex or wave. The results of our numerical experiments agree with this curve of baroclinic instability. If the situation is under baroclinic instability in one run, the perturbation with baroclinic instability has a tilted axis of the vortex and lee vortex in the middle layer are difficult to grow because that has a vertical axis of vortex. But if there is no baroclinic instability in the middle layer, lee vortex ceases to feel a 3-D tilted axis of vortex, so lee vortex can exist with a long lifetime. The run that is near the baroclinic instability curve is in critical situation that contains 3-D baroclinically vortices (or waves) and 2-D lee vortex, so it is difficult to interpret those vortices with its complexity.

5. Conclusion

To investigate the effects of the rotating stratified fluid system by the great mountain, several numerical experiments were executed with using 3-D non-hydrostatic meteorological prognostic model. We separated the layer as “top layer” and “middle layer” with Hunt et al. (1997 and 2001), and experiments for the generating of vortex were done.

In the top layer, Huppert and Bryan (1976) showed the initial generating vortex and its quasi-geostrophic theorem, so we investigated a more detail insight for the vorticity of this initial generating vortex by various experiments with some parameters with various values. Our results generally agreed with the improvement of Huppert and Bryan’s theorem within $0.3 < Fr < 0.9$ and one of the most important of this result is the vorticity of this vortex is not depend on Ro or Coriolis parameter.

On the other hand in the middle layer, to investigate the Coriolis effects on lee vortex, we detected the vortex formation in the middle layer and examined the relationship between the vortex formation and the characteristics with the run’s parameter as Ro and Fr. Differing from the non-rotating stratified system whose lee vortex is decided by Re and Fr, the formation of the vortex depended on both Fr and Ro in the rotating stratified system, so we assume the baroclinic instability and the scale of the perturbation and drew the critical line of the baroclinic instability. This line almost agreed with our detection of the vortex formation. This interpretation is that if the situation of baroclinic instability in leeward of the mountain, lee wave that has the tilted axis of the vortex (or wave) is dominant and the formation of lee vortex is difficult, on the other hand if the situation is not baroclinic instability state, lee wave is not dominant and lee vortex can exist in the middle layer. The existence of the vortex in the middle layer depends on Fr and moreover Ro.

These investigations mean when the jet event occurs, Fr and Ro increase around the huge mountain, and also mean if the jet event over the mountain, for example Tibet, Alps, and Rocky mountain is occur, the vortex generation in the top layer occurs and also the vertical wind shear comes to generate the lee vortex in the middle layer.
References


Presentations and Publications


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Counter-Rotating Quasigeostrophic Ellipsoidal Vortex Pair

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Abstract
In geophysical flows, coherent vortex structures persist for long time and their interactions dominate the dynamics of geophysical turbulence. Miyazaki et al. derived a Hamiltonian dynamical system describing the interactions of N ellipsoidal vortices, where each coherent vortex was modeled by an ellipsoid of uniform potential vorticity embedded in a locally uniform 3D shear field. In this paper, the interaction of counter-rotating vortex pair (dipole) was investigated. According to the ellipsoidal moment model, the inclination from the vertical axis approaches $\pi/2$ and one of the principal axes of the ellipsoid grows exponentially with time, if two slender vortices are placed within a critical distance initially. Direct numerical simulations, based on the Contour Advective Semi-Lagrangian algorithm, were performed in order to assess the validity of the model. The CASL computation tells that the vortices keep the dipole structure quite robustly. They survive from tilting down by emitting thin filaments from their top and bottom, even if they are placed very close.

Keywords : Quasigeostrophic vortices, CASL algorithm, Ellipsoidal Moment Model

1. Introduction
In geophysical flows, coherent vortex structures persist for long time and their interactions dominate the turbulence dynamics. Recently, a vortex-based turbulence model (so called ‘ellipsoidal moment model’: EMM) is constructed, where each coherent vortex is modeled by an ellipsoid of uniform potential vorticity (Miyazaki et al., 2001). In this paper, we investigate the motion of a counter-rotating pair of two vortices, called ‘dipole’. A stable dipole translates for a long distance and may play an important role in the ‘long range scalar transport’ in geophysical flows. There have been many works on the motion and stability of a two-dimensional dipole (Flierl, 1987), but we know little about the behavior of a vertically off-setted three-dimensional dipole.

2. Prediction of the Ellipsoidal Moment Model
We consider the motion of two interacting ellipsoidal vortices of uniform potential vorticity $\alpha_1=-\alpha_2=1$, whose centers of vorticity are located at $(X_{1,2}, Y_{1,2}, Z_{1,2})$. The potential vorticity is uniform inside both ellipsoids and the principal axis-lengths are denoted by $a_{1,2}, b_{1,2}, c_{1,2}$. Their orientations are specified by the Euler angles $\phi_{1,2}, \theta_{1,2}, \psi_{1,2}$. Their time evolution is governed by the ‘EMM equations’ (Miyazaki et al., 2001). The aspect ratios of vortices are taken to be $a_{1,2}/c_{1,2}=\beta_{1,2}/\gamma_{1,2}=0.3162$ and both spheroids are vertically standing at the initial time. The vortex heights are taken to be unity. The horizontal and vertical offset parameters are $a$ and $y$, respectively.

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Figure 1  Three patterns predicted by EMM

Figure 2  The inclination angle $\theta_1=\theta_2 : (a,h) = (0.8, 0.4)$, region (1).

We find three patterns depicted in Fig.1, i.e., (1) stable translation in the positive $y$-direction, (2) translation in the negative $y$-direction with large precessions and (3) singular behavior (tilting down) of both vortices. In region (1), the motion is doubly time-periodic with the periods 46.2 and 6.18. The translation velocity in the region (1) is always positive. The faster oscillation of smaller amplitude is superimposed on the main precession. The secondary oscillation is associated with the change of aspect ratio $\alpha_{1,2}/\beta_{1,2}$ in the plane perpendicular to the longest principal axis ($\gamma_{1,2}$). The inclination angle $\theta_1=\theta_2$ (Fig.2) and the
orientation angle $\varphi_1 = -\varphi_2$ oscillate quasi-periodically with small secondary faster oscillations being superimposed on the main oscillation.

Figure 3. The inclination angle $\theta_1 = \theta_2$ and the orientation angle $\varphi_1 = -\varphi_2$ for $a=0.6, h=0.95$: $(a, h) = (0.6, 0.95)$, region (2).

Figure 4. The inclination angle $\theta_1 = \theta_2$ and the orientation angle $\varphi_1 = -\varphi_2$: $(a, h) = (0.8, 0.4)$, region (3).
In region (2), both the inclination and orientation change non-periodically, as illustrated in Fig.3 (a=0.6, h=0.95). Untill t=160, the inclination angle increases and so does the longest principal semi-axis $\gamma_1$. For t>160, the largest axis $\gamma_1$ decreases and each ellipsoid takes a pancake-like form, i.e., $\alpha_1$ remains of order of unity and $\beta_1$ decreases to zero. The most striking thing occurs when the vertical off-set becomes smaller, i.e., as in region (3), both vortices are stretched infinitely. Because of this singular behavior, the EMM-computation stops in region (3), which is a serious drawback of EMM. The boundary-line between the regions (2) and (3) is about h/a=1.4. The orientation angle $\psi_1=-\psi_2$ tends to a certain limiting value (Fig.4).

3. CASL - Simulation

In order to investigate the effects of the deformation from an ellipsoid and the dissipative processes, (neglected in EMM), we perform direct numerical simulations of the quasigeostrophic equation based on the CASL-algorithm (Dritschel et al., 1997). The dissipative effects are taken into account by an artificial procedure called 'surgery'. A surgical operation is performed whenever the distance between two contours containing the same potential vorticity becomes less than a cut-off scale $\delta$. We have performed numerical computations corresponding to the cases in the previous section with the resolution $\Delta=2\pi/256$ ($128^3$ Fourier modes) and $\delta=\Delta/10$. Two runs with the initial conditions (a, h)=(1.2, 0.6) and (0.8, 0.4) are illustrated in Figs.5 and 6. These correspond to two of the qualitatively different cases shown in the previous section. (1) stable translation in the positive y-direction (Fig.5), (2) translation in the negative y-direction with large precessions: (not shown) and (3) tilting down of both vortices (a singular case): (a,h)=(0.8,0.4) (Fig.6). In case (1), both vortices precess slightly, retaining the ellipsoidal forms and the energy and enstrophy are conserved. In cases (2) and (3), the vortices tilt largely in the early stage of computation and emit filaments, later, from their top and bottom to nod up (dancing). In case (2), there remain many small ‘satellites’, after the vigorous filamentation. Similarly, filaments are emitted in case (3), but the remaining ‘satellites’ are dissipated later. The singular behavior predicted by EMM is not observed at all, being circumvented by the dissipative filamentation.

The behavior at later time is summarized in Fig.7, as a pattern map in the (a,h) plane. Open circles show the cases of stable translation without dissipation and crosses represent the cases of vortex-dancing with considerable filamentation. Squares represent the cases where many satellites remain after filamentation. The solid lines are thresholds predicted by EMM (see the previous section: Fig.1).

We can see that the model works practically well for vortices with small vertical off-sets (h<1), processes by giving alarms, i.e., by showing infinite stretching or large precession of one or both of the contour-rotating vortices. When the vortices are largely off-setted (h>1), non-ellipsoidal deformation becomes important, even if EMM predicts stable translation. Note that the dipoles never translate in the negative direction in the CASL-computations.
Figure 5  Time evolution of the stable case (1), \((a, h) = (0.8, 0.4)\).
Figure 6  Time evolution of the emitting filaments (with negligible satellites): case (3), (a, h) = (0.8, 0.4).
4. Summary

We have investigated the interaction of two counter-rotating vortices of the same strength (vortex dipole) based on the ellipsoidal moment model, which was extracted from the partial differential equations governing the dynamics of quasigeostrophic fluid motion. Direct numerical simulations (CASL) have been performed, in order to assess the validity (limitation) of the model prediction. The ellipsoidal moment model (EMM) tells that two vortices are stretched infinitely when they are placed within a certain critical distance initially. According to the CASL simulations, a dipole emits filaments vigorously, near the region where EMM predicts singular behavior. No really catastrophic behavior has been observed in the CASL simulations. The dipole structure remains robustly, even if the two counter-rotating vortices are almost in touch initially. The critical distance of the model does not coincide sharply with the results of the CASL simulations for largely off-settled cases. EMM captures only the tendency of the occurrence of dissipative events, by giving 'alarms', i.e., the infinite stretching of the contour-rotating vortices. EMM needs such refinements as proposed by Dritschel and his group, where an ellipsoid is represented by seven point vortices (McKiver et al., 2003).

References


Presentations and Publications

Reviewed Papers

Proceedings

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Development of a High-resolution Local Atmospheric Numerical Model in Cartesian Coordinate

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Abstract
A high-resolution local atmospheric numerical model is developed in Cartesian coordinate, and it is expected to suitably treat the steep topography and complex objects on the earth’s surface with a finer resolution. In this work, the finite volume method (FVM) in conjunction with the SIMPLER (Semi-Implicit Method for Pressure-Linked Equation Revised) algorithms is used for calculations of the unsteady, three-dimensional, compressible Navier-Stokes equations on a staggered grid. Abandoning the customary terrain-following normalization, we choose the Cartesian coordinate in which the height is used as the vertical one. Blocking-off method is introduced to handle all of the steep topography and complex objects above the earth’s sea-mean level. Here, the model has been run on calculating turbulent flow in an urban city by Large Eddy Simulation (LES). The test shows a satisfied result.

Keywords: High-resolution numerical model, Steep topography, Complex objects, Cartesian coordinate

1. Introduction

With the continuing increase in computational resources, the meso-scale models are being run at successively higher resolutions. It is reasonable to expect that a goal of running a meso-scale model at a horizontal resolution of O (100) m may be attainable in the near future, and the topography may then be more accurately represented. In such a situation, it seems natural to search for an alternative that will be better suited to handle the steep orography and complex objects for the high-resolution model simulations.

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

In this report, we present the advanced numerical methods based on finite volume discretization, i.e., blocking-off method for handling complex geometry, and incorporate it into coding of a robust, efficient and accurate dynamical core for the next-generation atmospheric meso-scale numerical model which is expected to suitably treat the steep topography and complex objects with a finer resolution for high-resolution meso-scale flow simulations.

2. Dynamical Core Descriptions

2.1 Governing Equations

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

Momentum equations:

\[
\frac{\partial \rho u}{\partial t} + \frac{\partial \rho uu}{\partial x} + \frac{\partial \rho uv}{\partial y} + \frac{\partial \rho uw}{\partial z} = - \frac{\partial p}{\partial x} + \rho f v + \mu \frac{\partial^2 u}{\partial x^2} + \mu \frac{\partial^2 u}{\partial y^2} + \mu \frac{\partial^2 u}{\partial z^2}
\]

\[
\frac{\partial \rho v}{\partial t} + \frac{\partial \rho vu}{\partial x} + \frac{\partial \rho vv}{\partial y} + \frac{\partial \rho vw}{\partial z} = - \frac{\partial p}{\partial y} - \rho f u + \mu \frac{\partial^2 v}{\partial x^2} + \mu \frac{\partial^2 v}{\partial y^2} + \mu \frac{\partial^2 v}{\partial z^2}
\]

\[
\frac{\partial \rho w}{\partial t} + \frac{\partial \rho uw}{\partial x} + \frac{\partial \rho vw}{\partial y} + \frac{\partial \rho ww}{\partial z} = - \frac{\partial p}{\partial z} - g + \mu \frac{\partial^2 w}{\partial x^2} + \mu \frac{\partial^2 w}{\partial y^2} + \mu \frac{\partial^2 w}{\partial z^2}
\]

Energy equation:

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

Continuity equation:

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

Equation of state for ideal gas:

\[P = \rho R T\]

2.2 Temporal and Spatial Discretizations

Detail of the numerical scheme on the temporal and spatial discretization used in the model is referred to Sha (2002).

2.3 Treatment of Irregularly Shaped Objective in Calculation Domain

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

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

3. Test Result

As test, the model has been run on a Large Eddy Simulation (LES) of turbulent flows in an urban city. Figure 1 shows a building block in an urban city and the grids set in the calculation. In Figure 2, we present the surface velocity in the building block. It illustrates the flow pattern characterized by the block. We found that no spurious flows are generated around the corners, and the simulation shows a satisfying result. Next work including three-dimensional computations on real flow in urban city is in progress.

![Figure 1](image)

Figure 1  Geometry and grid set in calculation.
Figure 2  Horizontal velocity $u$ near surface in building block.

Reference


Publications

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4. Other Research
Buoyancy Effects on Chemical Reaction in Thermally Stratified Liquid Turbulent Flows

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Abstract
The buoyancy effects on turbulent mixing and chemical reaction were numerically investigated using large-eddy simulation (LES) in two types of reacting liquid turbulence with density stratifications; a mixing layer and grid-generated turbulence. The present LES employed the dynamic SGS models for the SGS Reynolds stress and SGS mass and heat fluxes and our SGS model for the chemical reaction source term. The LES based on these SGS models was applied to our previous experiments on stratified reacting liquid turbulent flows. In order to examine the reliability of the present LES, the predictions of some turbulence statistics were compared with the measurements. The results show that the LES predictions are in good agreement with the measurements and the present LES can accurately estimate buoyancy effects on the diffusive-reactive mechanism in thermally stratified reacting liquid turbulent flows.

Keywords: Large-eddy simulation (LES), Chemical reaction, Liquid turbulence, Thermal stratifications

1. Introduction
Turbulent reacting liquid flows with density stratifications are often seen not only in industrial reactors but also in environmental water flows with chemical pollutants. It is, therefore, of great importance to investigate the reactive-diffusive mechanism in density-stratified reacting liquid turbulence. In spite of its importance, few numerical studies on the diffusive-reactive mechanism have been conducted in stratified liquid flows because of the high Schmidt number and complicated buoyancy effects.

One of the promising simulation techniques for such flows is the large-eddy simulation (LES) since direct numerical simulation (DNS) cannot be applied to practical reacting liquid flows with high Schmidt and Reynolds numbers. In the LES, large scale (grid scale: GS) quantities are directly computed, and small scale (subgrid scale: SGS) quantities are predicted by SGS models. In order to apply the LES to practical reacting flows with density stratifications, a SGS model which describes the diffusive-reactive mechanism and buoyancy effects is required. Gao and O’Brien (1993) have proposed one of the major SGS models. The model is based on the large eddy probability density function (LEPDF) of concentration of chemical species and the LEPDF can be derived from a transport equation of the LEPDF. However, the equation needs a lot of closure hypotheses, which results in a high computational cost. Another one is the β assumed-pdf approximation model by Cook and Riley (1994). However, the model is not applicable to liquid turbulence since the SGS concentration variance, which is important in liquid turbulence, has not been considered.

The previous SGS modeling has suggested that an effective SGS model, which is
applicable to reacting liquid turbulence, should be developed. Recently we have developed a simple but powerful SGS model based on the $\beta$-pdf model in which the SGS concentration variance is considered (Michioka et al., 2001) and we have examined the applicability and reliability of our proposed SGS model by comparing with the measurements in reacting liquid turbulence without density stratification. However, the applicability of our proposed SGS model to reacting liquid flow with stable and unstable density-stratifications has not been confirmed.

The purpose of this study is, therefore, to examine the applicability and reliability of our LES to density-stratified reacting liquid flows by comparing with our measurements (Onishi et al., 2003) in stratified conditions.

2. Large-Eddy Simulation

The governing equations are filtered continuity equation, momentum, mass and heat conservation equations:

$$\frac{\partial \overline{U_i}}{\partial t} + \frac{\partial \overline{U_j}}{\partial x_j} = 0,$$

$$\frac{\partial \overline{U_i}}{\partial t} + \overline{U_j} \frac{\partial \overline{U_i}}{\partial x_j} = -\frac{1}{\rho} \frac{\partial \overline{\rho}}{\partial x_i} + \nu \frac{\partial^2 \overline{U_i}}{\partial x_j \partial x_j} - \beta g_y (\overline{T} - T_s),$$

$$\frac{\partial \overline{C_i}}{\partial t} + \overline{U_j} \frac{\partial \overline{C_i}}{\partial x_j} = \frac{\nu}{\delta_{ij}} \frac{\partial^2 \overline{C_i}}{\partial x_j \partial x_j} - \frac{\partial q_{ij}}{\partial x_j} + \frac{\partial \omega_i}{\partial x_j},$$

$$\frac{\partial \overline{T}}{\partial t} + \overline{U_j} \frac{\partial \overline{T}}{\partial x_j} = \frac{\nu}{\delta_{ij}} \frac{\partial^2 \overline{T}}{\partial x_j \partial x_j} - \frac{\partial h_i}{\partial x_j},$$

where an overbar indicates the filtered value. The effects of the unresolved SGS in the equations (2) to (4) appear in the SGS stress term $\tau_{ij}$, SGS mass flux term $q_{ij}$, SGS heat flux term $h_i$, and filtered chemical reaction source term. These terms need to be modeled to represent the SGS effects by the filtered quantities at the resolved scales (GS). In this study, dynamic SGS models were employed to the SGS Reynolds stress and SGS mass and heat fluxes. For the filtered reaction source term, the SGS model developed in our previous study (Michioka et al., 2001) was used.

2.1 Modeling for the Filtered Reaction Source Term

A second-order rapid reaction ($A + B \rightarrow P$) between acetic acid and ammonium hydroxide was used here. In such a rapid reaction, the timescale of the chemical reaction is far smaller than that of the turbulent diffusion. This means that the time step $\Delta t$ must be sufficiently smaller than the time scale of chemical reaction. However, such an extremely small time step cannot be treated even by a supercomputer. Therefore, the SGS model of Cook and Riley (1994) was modified to be applicable to a turbulent liquid flow with a rapid reaction. The model introduces a conserved scalar $\zeta$ and needs the SGS fluctuation of $\zeta$. In liquid flows, the SGS fluctuation is not equal to the GS fluctuation because of its high $Sc$. To consider the SGS effect, the SGS fluctuation was assumed to be correlated with a test-scale fluctuation as

$$\overline{\zeta}^2 \approx c_f \overline{\zeta}^2$$
where \( \sim \) indicates the test-scale filtered value. The scale of the test filter was twice of the normal SGS filter. The correlation coefficient of \( c_{ij} = 5.0 \) was obtained from the DNS of stationary isotropic liquid turbulence (Michioka et al., 2001).

2.2 Modeling of Buoyancy Effect on SGS Flow Field

In a conventional dynamic SGS model for the SGS Reynolds stress, energy generation by shear is assumed to be equal to energy dissipation. From this assumption, we can derive

\[
\tau_y - \frac{1}{3} \delta_y \tau_{xx} = -2c^2 \Delta^{-2} |S| \overline{S_y}.
\]

(6)

In the above equation, \( \overline{S_y} = \frac{\partial u_i}{\partial x_i} + \frac{\partial u_j}{\partial x_j} \)/2, and the coefficient \( c \) can be determined dynamically from GS quantities.

In stratified flows, the buoyancy may directly affect the SGS flow field. Especially in liquid flows, the buoyancy effect seems to be significant because of high \( Pr \). Wong and Lilly (1994) assumed that energy generation by shear and buoyancy is balanced with energy dissipation, and they derived

\[
\tau_y - \frac{1}{3} \delta_y \tau_{xx} = -2c_{sGS} \overline{S_y}
\]

\[
= -2c^2 \Delta^{-2} \left( 1 + \frac{g_{\beta}}{Pr_{sGS} |S|} \frac{\partial \Delta}{\partial S} \right) |S| \overline{S_y}.
\]

(7)

By comparing the two LES predictions based on the above equations (6) and (7), the buoyancy effect on the SGS flow field will be discussed.

2.3 Computational Method

The governing equations for the LES are given by equations (1) to (4). These equations were discretized on the staggered mesh arrangement to construct a finite-difference formation. The HSMAC method was used to solve the discretized equations. The second-order central difference method and the second-order Runge-Kutta method were used to calculate the spatial derivatives and the time integration, respectively.

The LES was carried out for two types of reacting liquid turbulent flows with thermal stratifications. One was a typical turbulent shear flow; a mixing layer, and the other was a typical isotropic turbulent flow; grid-generated turbulence.

2.3.1 Mixing Layer

Figure 1 shows the schematic diagram of the computational domain for the mixing layer. In the neutral and stable cases, the computational domain was \( 480 \times 80 \times 80 \) mm in the streamwise, vertical, and spanwise directions and the numbers of grid points were \( 360 \times 60 \times 60 \). In the unstable case, the mixing region was enlarged vertically, so that the domain was set to \( 480 \times 106 \times 80 \) mm and the number of grid points was set to \( 360 \times 80 \times 60 \).

Table 1 shows the computational conditions for a mixing layer. The averaged streamwise velocity was set to 0.125 m/s. The upper-layer streamwise velocity was set to 0.08 m/s larger than the lower-layer one. In the unstable case, the upper-layer temperature was set to 10 K lower than the lower-layer one. In the stable case, the upper-layer temperature was set to 10 K higher than the lower-layer one.
Table 1 Computational conditions for a liquid mixing layer.

<table>
<thead>
<tr>
<th></th>
<th>$U_0$ [m/s]</th>
<th>$\Delta U$ [m/s]</th>
<th>$\Delta T$ [K]</th>
</tr>
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<tbody>
<tr>
<td>neutral case</td>
<td>0.125</td>
<td>0.080</td>
<td>0</td>
</tr>
<tr>
<td>unstable case</td>
<td>0.125</td>
<td>0.080</td>
<td>-10</td>
</tr>
<tr>
<td>stable case</td>
<td>0.125</td>
<td>0.080</td>
<td>10</td>
</tr>
</tbody>
</table>

Figure 1  The computational domain for a mixing layer.

2.3.2 Grid-Generated Turbulence

Figure 2 shows the schematic diagram of the computational domain for the grid-generated turbulence. The computational domain was $520 \times 80 \times 80$ mm and the number of grid points was $260 \times 80 \times 80$. The mesh size of the turbulence-generating grid, $M$, and the thickness of the square rod, $d_t$, were $2.0 \times 10^{-2}$ m and $2.0 \times 10^{-3}$ m, respectively. Table 2 shows the computational conditions for the grid-generated turbulence. The streamwise velocity was set to $0.180$ m/s. Under these conditions, the Reynolds number based on the mesh size ($Re_M$) was $4.0 \times 10^3$. Bulk Richardson numbers ($Ri_b$) were $1.4 \times 10^2$ and $-1.4 \times 10^{-2}$ in stable and unstable cases, respectively.

Table 2 Computational conditions for grid-generated turbulence.

<table>
<thead>
<tr>
<th></th>
<th>$U_0$ [m/s]</th>
<th>$\Delta T$ [K]</th>
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<tr>
<td>stable case</td>
<td>0.180</td>
<td>10</td>
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</table>
3. Results and Discussions

3.1 Mixing Layer

Figure 3 shows the streamwise distributions of the mixing layer width $\delta_u$ in the velocity field. Two LES predictions are shown in the figure together with the measurements. Hereafter we refer the two LESs based on the equations (6) and (7) without and with buoyancy effect as LES(1) and LES(2), respectively. The both LES predictions are in good agreement with the measurements. In the neutral case, the mixing layer width grows linearly in the region of $x > 0.15 \, \text{m}$. This means that in neutral stratification a typical mixing layer is formed in the present LES. The value of $\delta_u$ decreases in stable stratification, whereas $\delta_u$ increases in unstable stratification. Furthermore, it is found that in unstable and stable stratifications there is little difference in $\delta_u$ between the two predictions by LES(1) and LES(2). The little difference was also seen in other turbulent quantities. This means that the buoyancy does not affect directly on the SGS flow field and the SGS Reynolds stress (equation (6)) without buoyancy effect is applicable even to stratified liquid mixing layer and grid-generated turbulent flows. Therefore, the predictions only by LES(1) will be shown in the following figures.

Figure 4 shows the vertical distributions of the turbulence intensities of vertical velocity fluctuation at $x = 0.40 \, \text{m}$. Here $y_v$ is the vertical position with the mean velocity equivalent to the cross-sectional mean velocity. The promotion and suppression of turbulence can be seen in unstable and stable stratifications, respectively. The LES predictions are in good agreement with the measurements and the turbulence promotion and suppression by buoyancy are well predicted. These agreements between the LES predictions and the measurements in figures 3 and 4 show that the present LES can accurately predict the buoyancy effects on the velocity field.

Figure 5 shows the vertical distributions of the vertical mass flux at $x = 0.40 \, \text{m}$. Here $y_m$ is defined by the vertical position with the mean concentration of species A equivalent to the cross-sectional mean concentration of species A. The mass transfer is promoted by unstable stratification and suppressed by stable stratification. The present LES can well estimate the buoyancy effects on the turbulent mass transfer. Figure 6 shows the vertical distributions of the mean concentration of chemical product at $x = 0.40 \, \text{m}$. The chemical reaction is also promoted by unstable stratification and is suppressed by stable stratification. Although there is a little difference between the predictions and measurements, the present LES well explains
the measurements. These results show that the present LES can accurately estimate the buoyancy effects on the diffusive-reactive mechanism in the thermally stratified mixing layer.

Figure 3  Streamwise distributions of the mixing layer width in the velocity field in a mixing layer.

Figure 4  Vertical distributions of the turbulence intensities of vertical velocity fluctuation at \( x=0.40 \) \( \text{m} \) in a mixing layer.
Figure 5  Vertical distributions of the vertical mass flux at $x=0.40$ m in a mixing layer.

Figure 6  Vertical distributions of the mean concentration of chemical product at $x=0.40$ m in a mixing layer.
3.2 Grid-Generated Turbulence

Figure 7 shows the vertical distributions of the turbulence intensities of vertical velocity fluctuation on the centerline \((y/M = z/M = 0)\) in grid-generated turbulence. The intensity decays exponentially in neutral stratification and the exponent of -1.6 was close to the value of -1.4 obtained by Stapountzis et al. (1986). Although there is a little difference between the predictions and measurements in the upstream region \(x/M < 10\), the agreement is good in the downstream region. This means that in neutral stratification a typical isotropic grid-generated turbulence is formed in the present LES. In addition, the turbulence promotion and suppression by unstable and stable stratifications are well predicted. These results show that the present LES can well predict the thermally stratified flow fields not only in the mixing layer but also in the grid-generated turbulence.

Figure 8 shows the vertical distributions of the vertical mass flux at \(x/M=20\). The mass transfer is strongly affected by thermal stratifications as well as in the mixing layer. The present LES can well predict the buoyancy effects on the turbulent mass transfer in thermally stratified grid-generated turbulence.

Figure 9 shows the vertical distributions of the mean concentration of chemical product at \(x/M=20\). The chemical reaction is promoted and suppressed by unstable and stable stratifications, respectively. The predictions are in good agreement with the measurements. These results also show that the present LES can accurately predict the buoyancy effects on the diffusive-reactive mechanism not only in the thermally stratified mixing layer but also in the thermally stratified grid-generated turbulence.

![Figure 7](image-url)  
**Figure 7** Streamwise distributions of the turbulence intensities of vertical velocity fluctuation in grid-generated turbulence.
Figure 8  Vertical distributions of the vertical mass flux at $x/M=20$ in grid-generated turbulence.

Figure 9  Vertical distributions of the mean concentration of chemical product at $x/M=20$ in grid-generated turbulence.
4. Conclusions

The present LES based on the SGS model proposed for the filtered chemical reaction source term was applied to two types of thermally stratified reacting liquid turbulence; a mixing layer and grid-generated turbulence. The results from this study can be summarized as follows.

1. The buoyancy effects induced by unstable and stable thermal stratifications on the SGS flow field are negligibly small even in liquid mixing layer and grid-generated turbulence.
2. The present LES can well predict the buoyancy effects and diffusive-reactive mechanism in thermally stratified reacting liquid turbulence.

References


Publications and Presentations

**Original Papers and Reviews:**


**Conference Reports:**


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Simulation of Drying Phenomena Associated with Vegetation Change by Using NICE Model in the Kushiro Mire

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Abstract
We developed a NIES Integrated Catchment-based Eco-hydrology (NICE) model. New process-based model which includes surface-unsaturated–saturated water processes and land-surface processes describing the variation in phenology assimilated with MODIS satellite data was developed. This model was applied to the Kushiro River catchment (northern Japan, area: 2204.7 km²) with a resolution of 500 m and 8 days averaged vegetation changes, and compared with field data for soil moisture, soil temperature, groundwater level, and river flow discharge during two years. Excellent agreement between simulated and measured values was obtained, showing that this model achieves a high accuracy by taking into account vegetation phenology, soil property and geological structure. This model explains water cycle change and drying phenomena in the Kushiro mire caused by the interaction between the increased sediment load from river channelization and the invasion of *Alnus japonica* into the mire.

Keywords: Drying phenomena, Kushiro mire, MODIS satellite data, Integrated Catchment-based Eco-hydrology model

1. Introduction

The Kushiro River catchment (area: 2204.7 km² located in northern Japan), shown in Figure 1 (the Digital National Land Information GIS data of Japan, 1993), consists of a low elevation area, less than 10 m at the downstream regions, and moderately high mountains in 65 % of the upstream regions, with the highest point more than 1000 m above sea level. Most of the area is covered by forest or farmland, and at its lower end lays the Kushiro mire, the largest mire in Japan, which was listed under the Ramsar Convention in 1980 (mean elevation: 2–10 m, area: 176.74 km²). Due to the channelization, runoff containing nutrients from farmland and sediments from short-cut channels (arrow) have inflowed directly into the mire and deposited flood-borne sediment. Sediment deposition in the mire has caused topographical changes, lowering the groundwater level, which has resulted in the soil starting to dry out (Environment Agency of Japan, 1984, 1993). Consequently, *Alnus japonica*, belonging to the class of *Betulaceae*, a deciduous tree with a height of approximately 15 m has invaded the mire in the downstream areas of the Kucyoro River and has increased its distribution. *Alnus japonica* has propagated widely around the Kushiro mire after channelization along the downstream end of the Kucyoro River, owed mainly to the lowering of the groundwater level and the increased nutrient input, resulting in the mire gradually shrinking. This shows that human activities have changed the water cycle in the Kushiro mire and the change in the water cycle caused the vegetation succession. With these conditions,
there is an urgent need to understand the water cycle and soil moisture dynamics in the whole Kushiro River catchment, including the mire, for the protection of the Kushiro mire (Environment Agency of Japan, 1984, 1993).

The objective of the current research is to simulate the drying phenomena in the Kushiro mire due to the effects of ecosystem changes. Thus, a new integrated catchment model is indispensable to evaluating the Kushiro River catchment. In particular, land-surface processes including the variations in phenology, and surface unsaturated-saturated water processes, must be included to evaluate the catchment’s water cycle changes. In the present study, we developed an NIES Integrated Catchment-based Eco-hydrology (NICE) model, which reproduces water cycle changes and drying phenomena in the Kushiro mire caused by the interaction between the increased sediment load, due to river channelization, and the invasion of *Alnus japonica* into the mire.

### 2. Model Description

#### 2.1 General Model Framework

We developed NIES Integrated Catchment-based Eco-hydrology (NICE) model (Nakayama and Watanabe, 2003), consisted of SiB2 (Sellers, *et al.*, 1996) for soil moisture and heat flux, the USGS MODFLOW model of three-dimensional groundwater flow (McDonalds and Harbaugh, 1988), and a grid-based hydrology model (Figure 2). MODIS satellite data with a 1 km mesh were inputted to the NICE model for describing the spatial and temporal changes of vegetation phenology. The water flux between recharge layer and groundwater layer was calculated in order to combine soil moisture model and groundwater flow model in each time step. Furthermore, the effective precipitation and the seepage between river and groundwater are included in NICE model. Therefore, the NICE model can
reproduce long-term components of river flow discharge due to recharge rates in addition to short-term components.

2.2 Biophysical and Soil Moisture Models

Since a detailed description of SiB2 can be found in a previous study (Sellers, et al., 1996), only a brief description of heat and water transfer is given here. SiB2 performs numerical simulations of land–atmosphere interactions based on the principals of energy and water conservation, including ecological processes such as vegetation phenology. SiB2 divides canopy into two layers (canopy layer and ground surface), and soil layer into three layers (upper layer, intermediate layer, and lower layer) in a vertical one dimension. The governing equations for SiB2 prognostic variables consists of temperatures, interception stores, and soil moisture stores.

(a) Canopy, ground surface, and deep soil temperatures

\[
C_c \frac{\partial T_c}{\partial t} = Rn_c - H_c - \lambda E_c - \xi_{cs},
\]

\[
C_g \frac{\partial T_g}{\partial t} = Rn_g - H_g - \lambda E_g - \frac{2\pi C_d}{\tau_d} (T_g - T_d) - \xi_{gs},
\]

\[
C_d \frac{\partial T_d}{\partial t} = \frac{1}{2\sqrt{365\pi}} (Rn_g - H_g - \lambda E_g)
\]

where the subscript “c” refers to the canopy, “g” to the soil surface, “d” to the deep soil. \(T_c\), \(T_g\), and \(T_d\) (K) are canopy, ground surface, and deep soil temperatures, \(Rn_c\) and \(Rn_g\) (W/m²) are absorbed net radiation of canopy and ground, \(H_c\) and \(H_g\) (W/m²) are sensible heat flux, \(E_c\) and \(E_g\) (kg/m²/s) are evapotranspiration rates, \(C_c\), \(C_g\), and \(C_d\) (J/m²/K) are effective heat capacities, \(\lambda\) (J/kg) is latent heat of vaporization, \(\tau_d\) (s) is daylength, and \(\xi_{cs}\) and \(\xi_{gs}\) (W/m²) are energy transfer due to phase changes in \(M_c\) and \(M_g\).
(b) Interception stores

\[
\frac{\partial M_c}{\partial t} = P - D_d - D_c - E_{ci} / \rho_w \tag{4}
\]

\[
\frac{\partial M_g}{\partial t} = D_d + D_c - E_{gi} / \rho_w \tag{5}
\]

where \( M_c \) and \( M_g \) (m) are water or snow-ice stored on the canopy and on the ground, \( P \) (m/s) is precipitation rate, \( D_d \) (m/s) is canopy throughfall rate, \( D_c \) (m/s) is canopy drainage rate, \( E_{ci} \) and \( E_{gi} \) (kg / m² / s) are interception loss of canopy and ground, and \( \rho_w \) (kg/m³) is density of water.

(c) Soil moisture stores

\[
\frac{\partial W_i}{\partial t} = \frac{1}{\theta_i D_i} [P_{wi} - Q_{i,2} - \frac{1}{\rho_w} E_{gi}] \tag{6}
\]

\[
\frac{\partial W_j}{\partial t} = \frac{1}{\theta_j D_j} [Q_{j,2} - Q_{j,3} - \frac{1}{\rho_w} E_{cj}] \tag{7}
\]

\[
\frac{\partial W_k}{\partial t} = \frac{1}{\theta_k D_k} [Q_{k,3} - Q_k] \tag{8}
\]

where \( W_i \) is soil moisture fraction of \( i\)-th layer (=\( \theta / \theta_i \)), \( \theta_i \) (m³/m³) is volumetric soil moisture in \( i\)-th layer, \( \theta_i \) (m³/m³) is value of \( \theta \) at saturation, \( D_i \) (m) is thickness of the soil layer, \( Q_{ij} \) (m/s) is the flow between \( i \) and \( j \) layers, \( Q_s \) (m/s) is gravitational drainage from recharge soil moisture store, \( E_{ci} \) (kg / m² / s) is canopy transpiration, \( E_{gi} \) (kg / m² / s) is ground evaporation, and \( P_{wi} \) (m/s) is infiltration of precipitation into the upper soil moisture store.

2.3 Groundwater Model

MODFLOW numerically solves a three-dimensional groundwater flow equation for a porous medium by using a finite-difference method to simulate the distribution of hydraulic head within the groundwater (McDonalds and Harbaugh, 1988). A partial-differential equation of three-dimensional groundwater flow is expressed in the following equation.

\[
\frac{\partial}{\partial x}\left(K_{xx} \frac{\partial h_x}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy} \frac{\partial h_y}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz} \frac{\partial h_z}{\partial z}\right) + W = S_x \frac{\partial h_x}{\partial t} \tag{9}
\]

where \( K_{xx}, K_{yy}, \) and \( K_{zz} \) (m/s) are values of hydraulic conductivity along the \( x, y, \) and \( z \) coordinate axes, \( h_x \) (m) is the potentiometric head, \( W \) (1/s) is the volumetric flux per unit volume representing sources and/or sinks of water, \( S_x \) (1/m) is the specific storage, respectively. The above-mentioned equation (9) is discretized by using the finite-difference method on the assumption that the variables between two cells change linearly.

\[
CR_{i,j-\frac{1}{2},k} \left(h_{i,j-\frac{1}{2},k} - h_{i,j,k}\right) + CR_{i,j+\frac{1}{2},k} \left(h_{i,j+\frac{1}{2},k} - h_{i,j,k}\right) + CC_{i,j-\frac{1}{2},k} \left(h_{i,j-\frac{1}{2},k} - h_{i,k}\right) + CC_{i,j+\frac{1}{2},k} \left(h_{i,j+\frac{1}{2},k} - h_{i,k}\right) + CV_{i,j-\frac{1}{2},k} \left(h_{i,j-\frac{1}{2},k} - h_{i,j-1,k}\right) + CV_{i,j+\frac{1}{2},k} \left(h_{i,j+\frac{1}{2},k} - h_{i,j+1,k}\right)
\]

\[
+ P_{i,j,k} h_{i,j,k} + Q_{i,j,k} = SS_{i,j,k} \left(DELR \times DELC \times THICK \times i \times DELR \times DELC \times THICK \times i \right) \frac{h_{i,j,k} - h_{i,j,k-1}}{i^{m-1}} \tag{10}
\]

where \( h_{i,j,k} \) is head at cell \( i,j,k \) at time step \( m \), \( CV, CR, \) and \( CC \) are hydraulic conductances or branch conductances between node \( i,j,k \) and a neighboring node, \( P_{i,k} \) is the sum of coefficients of head from source and sink terms, \( Q_{i,j,k} \) is the sum of constants from source and sink terms with \( Q_{i,j,k} < 0.0 \) for flow out of the groundwater system and \( Q_{i,j,k} > 0.0 \) for flow in,
\( \text{DELR}_i \) is the cell width of column \( j \) in all rows, \( \text{DELC}_i \) is the cell width of row \( i \) in all columns, \( \text{THICK}_{i,j,k} \) is the vertical thickness of cell \( i,j,k \), and \( r^m \) is the time at time step \( m \).

### 2.4 Surface Hydrology Model

The surface hydrology model consists of a hill-slope hydrology model and a distributed stream network model based on both a kinematic and a dynamic wave theories. The hill-slope hydrology model consists of a water and thermal energy budget model and a surface runoff model. The kinematic wave model for distributed surface runoff can be expressed by the following equations (Takasao and Shiiba, 1988):

\[
\frac{\partial h}{\partial t} + \frac{1}{b(x)} \frac{\partial}{\partial x} \left\{ q b(x) \right\} = r(x,t) \cos \theta(x) \tag{11}
\]

\[
q = \frac{k \sin \theta(x)}{\gamma} h_a, \quad (0 < h_a < d),
\]

\[
q = \sqrt{\frac{\sin \theta(x)}{n} (h_a - d)^m + \frac{k \sin \theta(x)}{\gamma} h_a}, \quad (h_a \geq d)
\tag{12}
\]

where \( q \) (m²/s) is the discharge of unit width, \( r(x,t) \) (m/s) is the effective rainfall intensity at position \( x \) and time \( t \), \( b(x) \) (m) is the width of the flow, \( \theta(x) \) is the river bed gradient, \( k \) (m/s) is the hydraulic conductivity in a “A-layer” with a depth of \( D \) (m) near the ground surface, \( n \) (m/s) is the Manning coefficient, and \( m = 5/3 \). When \( H \) (m) is defined as the depth of the rainwater flow in the A-layer, \( h_a \) (m) as the apparent water depth (= \( \gamma H \)), and \( \gamma \) as the porosity of the A-layer, and \( d = cD \), then the true water depth is given by \( h_a/\gamma \) for \( h_a < d \), and by \( d/\gamma + h_a - d \) for \( h_a \geq d \).

The basic equation of one-dimensional unsteady flow is expressed in the following continuity equation (13) and momentum equation (14).

\[
\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q_i \tag{13}
\]

\[
\frac{\partial Q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{Q^2}{A} \right) + gA \frac{\partial h}{\partial x} + gA(I_f - i) = 0 \tag{14}
\]

where \( A \) (m²) is the cross-section area, \( Q \) (m³/s) is the discharge, \( q_i \) (m³/s) is the lateral inflow \( q \) entering along the side of the river channel simulated by a hill slope model, \( h_i \) (m) is the water depth in river, \( g \) (m/s²) is the gravitational acceleration, \( I_f \) is the friction slope, and \( i \) is the bed slope, respectively. The first term on the left-hand side of equation (14) is the local acceleration term, the second is the convective acceleration term, the third is the pressure force term, and the forth is the friction force term, and the fifth is the gravity force term, respectively. The local and convective acceleration terms represent the effect of inertial forces on the flow. The alternative distributed flow routing models are produced by using the full continuity equation while eliminating some terms of the equation (14). The simplest distributed model is the kinematic wave model, which neglects the local acceleration, convective acceleration, and pressure terms in the equation (14); that is, it assumes \( I_f = i \) and the friction and gravity forces balance each other. The diffusion wave model neglects the local and convective acceleration terms but incorporates the pressure term. The dynamic wave model consists all the acceleration and pressure terms in the equation (14). The equations (13) and (14) are discretized by using Preissmann’s implicit scheme in the non-linear simultaneous equations, and solved by using the Newton–Raphson iteration method for convergence and correcting the approximate solution. Finally, the discharge and water depth at each point are given.
In this study, the kinematic wave theory was applied to hill-slope runoff at each cell of the 50 m mesh by using a DEM of 50 m mesh (Geographical Survey Institute of Japan, 1981) all through the Kushiro River catchment in Figure 1. Then, both the kinematic and dynamic wave theories were applied to a stream network of 317 rivers by inputting the discharges at the upper river channels and the lateral flows by hill slope model. The dynamic effects are expected to be more important where stream slopes generally do not exceed 3/1000 (Samuels and Skeels, 1990; Meselhe and Holly, 1997). So, dynamic wave theory was applied for simulating large flat area around the Kushiro mire.

2.5 Integration of Models

In order to combine unsaturated flow and saturated flow, the water flux \( q_f \) is expressed by using the gradient of hydraulic potentials between the deepest layer of unsaturated flow and the groundwater level in the expansion of the drainage in the original SiB2 model (Sellers et al., 1996) in the following equation (15).

\[
q_f = -K \nabla \Psi = -K \frac{\Delta \Psi}{\Delta z} = -\frac{K}{D_s/2 + (D_s - h_g)} \left( \frac{\Psi_g - \Psi_3}{D_s/2 + (D_s - h_g)} + 1 \right) \tag{15}
\]

where \( K \) (m/s) is estimated effective hydraulic conductivity between unsaturated and saturated layers, \( \Psi_g = h_g \) and \( \Psi_3 = \psi_p + D_p + D_s/2 \) are hydraulic potentials both at groundwater surface and lower layer of unsaturated flow, \( D_s \) (m) is the distance between the top of the second layer and the bottom of the 20th layer in groundwater model, and \( h_g \) (m) is the hydraulic head simulated by groundwater model, respectively. When the groundwater level rises up and enters into the soil moisture layer, the partial pressure is set at the bottom of unsaturated layer \( \Psi_3 = \psi_p \) for simulating soil moisture. After the water flux \( q_f \) is calculated in each time step, the flows between each unsaturated soil layer \( Q_{ij} \) are simulated by using an improved backward-implicit scheme in order to simulate soil moisture \( \theta_i \) in \( i \)-th layer in the equations (6)-(8). Furthermore, this flux is inputted to the groundwater flow model as recharge rate at the upper boundary condition, and the groundwater flow model is simulated. For the treatment of effective rainfall intensity \( r \) of surface hydrology model, the effective precipitation can be calculated from the precipitation rate \( P \), the infiltration of precipitation into the upper soil moisture store \( P_{sat} \), and the evapotranspiration rates \( (E_v + E_s) \). The seepage between river and groundwater was also included by an application of Darcy’s Law (McDonalds and Harbaugh, 1988).

3. Observed Data and Boundary Conditions

Spatially distributed data sets with a resolution of 500 m were prepared for the simulations. To describe vegetation phenoology over space and time in the simulation, MODIS satellite data (1 km mesh) from 2001 to 2002 were used, including variations in \( FPAR \) and \( LAI \), after validation and verification of the MODIS data by ground-truth data collected at the Tomakomai Flux Tower, Hokkaido. Furthermore, three meteorological stations, 30 groundwater level meters, and 13 flow depth meters were set throughout the overall catchment, as shown in Figure 1, for the calibration and validation of numerical results.

3.1 Meteorological Data

Three meteorological stations were established in vegetation typical of the Kushiro River catchment, as shown in Figure 1 (mire: 43°06’05”N, 144°20’29”E, mean elevation 8 m;
grassland: 43°31’08”N, 144°28’10”E, mean elevation 187 m; forest: 43°20’18”N, 144°38’55”E, mean elevation 127 m). Meteorological variables were automatically recorded hourly at each station onto a computer. The data were collected from 2001 to 2002. The variables recorded (and instruments used) were air temperature, humidity, wind speed, net radiation, albedo, precipitation, ground temperature, soil moisture, and groundwater level.

Furthermore, measurements of the surface temperature at a 15 m height from the ground surface, FPAR at 25 and 40 m heights, and net radiation at a 25 m height taken from the Tomakomai Flux Tower (42°44’13.1”N, 141°31’7.1”E, mean elevation 115–140 m), which stands in a coniferous forest with a height of about 15-20 m, mainly larch (Larix), were provided by the Center for Global Environmental Research (CGER), NIES, to validate MODIS data against ground-truth data. In order to simulate the overall Kushiro River catchment, 11 points of AMeDAS (Automated Meteorological Data Acquisition System) data of hourly precipitation, air temperature, wind speed, and actual sunshine duration in the catchment collected by the Japan Meteorological Business Support Center were also used, in addition to three meteorological stations data of NIES.

3.2 Vegetations and Soil Properties

Vegetation class and soil texture data were also used for simulation under boundary conditions. Figure 3 shows the soil texture data digitized from a soil map of arable land in Hokkaido (Hokkaido National Agricultural Experiment Station, 1985). Soil texture data were categorized into 7 types for SiB2, and were converted into a 1 km mesh. The Kushiro mire mainly consists of organic soil, mainly peat. Vegetation class data categorized into 11 types, identified by the Ministry of Environment in 1993, were also converted into a 1 km mesh. The higher value depends on natural vegetation, showing that the Kushiro mire has the richest vegetation. This vegetation class was directly related to vegetation type for input parameter of SiB2. Furthermore, geological structure was divided into four types on the basis of hydraulic

Figure 3: Example of boundary conditions for simulation at 1 km mesh in the Kushiro River catchment, soil texture (1981).
conductivity, the specific storage of porous material, and specific yield to facilitate computational efficiency and numerical stability by using the explanatory text of the hydrogeological maps of Hokkaido (Ohata et al., 1975) for about 140 points.

3.3 MODIS Data

The MODIS-LST (Land Surface Temperature) data were converted to surface temperatures using the NASA MOD11-ATBD (algorithm theoretical basis documents) equation (Wan, 1999). \textit{FPAR} and \textit{LAI} were calculated using the MOD15-ATBD from the surface reflectance product (MOD09) and the land cover type product (MOD12), where \textit{FPAR} (\textmu mol/m\textsuperscript{2}/s) is fraction of photosynthetically active radiation, \textit{LAI} is leaf area index, and \textit{DN\textsubscript{i}} (i = 1, 2, 3) is the digital number of MODIS data for \textit{T\textsubscript{0}}, \textit{FPAR}, and \textit{LAI}, respectively. Because MODIS has more data channels than other previous satellites, such as Landsat, it is possible to analyze higher-order products, such as \textit{LAI} and \textit{FPAR}, which are important parameters for evaluating vegetation growth (Christopher et al., 1998). The MODIS data were averaged over each 3 \times 3 square of pixels including the effect of transformation error from sinusoidal to universal transverse Mercator coordinate system (UTM) after noise and null values due to bad weather were eliminated. MODIS data were compared with the synchronized ground-truth data at the Tomakomai Flux Tower. The surface temperature estimate is highly accurate (\textit{R\textsubscript{s}} = 0.992). Although the MODIS \textit{FPAR} value underestimates the observed value in winter (mainly because of higher cloud cover and falling snow), the correlation is still good (\textit{R\textsubscript{s}} = 0.788).

Figures 4a and 4b show the seasonal variations in MODIS \textit{LAI} and \textit{FPAR} (\textmu mol/m\textsuperscript{2}/s) images of the Kushiro River catchment in the non-snow period (1 May to 31 October) of 2001 (1 km mesh, image areas: 43°00’-45°N, 144°00’-145°E) with the vertical scale showing the units.

![Figure 4](image)

\textbf{Figure 4} Seasonal variation in MODIS LAI and FPAR images at 1km mesh (image areas: 43°00’-45°N, 144°00’-145°E). (a) LAI-Image from MODIS-Data. (b) FPAR-Image from MODIS-Data. Blue line is border of catchment, and pink line is that of the Kushiro mire.
Both values were obtained every 8 days throughout 2001 for input into the NICE model. In Figures 4(a, b), the blue line is the border of the catchment, and the pink line is that of the Kushiro mire. These figures show seasonal changes in $LAI$ and $FPAR$, particularly in the mire, implying that the vegetation phenology and the water cycle are closely related in the mire. Both parameters take their maximum values (green) from early summer to fall while vegetation is growing, and decrease toward winter, and both variables are highly correlated ($R_t = 0.891$). In order to apply both $LAI$ and $FPAR$ from MODIS data to the integrated model, a moving average was calculated after noise was removed. Because MODIS data are collected every 8 days, a constant value was input into the model for each block of 8 days in the simulation.

### 3.4 Groundwater and River Flow Data

Changes in groundwater levels at 30 sites and river flow-depths at 13 sites were measured every hour during the same period as the meteorological observations (Figure 1). The water level was automatically recorded in data-loggers every hour.

$H-Q$ curves (usually, with almost a one-to-one relationship between water level and stream discharge) were used to convert water level into river discharge, for flow-depth measurements in rivers. The relationship was poor in the mire because beds and cross-sections were frequently variable in rivers around the mire, each river flowing into the mire has its own runoff hydrography, and bed slopes are very small and are thus affected by downstream flow depths and floods.

### 4. Results and Discussions

The simulation area is 50 km wide by 80 km long, covering the whole Kushiro River catchment (Figure 1). This area is discretized into a grid of $100 \times 160$ blocks, with a grid spacing of 500 m. The simulation was conducted on an NEC SX-6 supercomputer at the CGER of NIES, Tsukuba. Simulations were performed for two years from 2001 to 2002, using the first 6 months as a warm-up period to reach equilibrium conditions, to conduct a parameter estimation, and to compare the steady-state values of the simulation with the observed values of previous. Hourly time steps were used for this simulation. The hydraulic conductivity for each cell was calibrated by trial and error on the basis of the measured values. The calibration consisted of fitting simulated hydraulic heads to the observed heads while keeping the hydraulic conductivity values in the initial estimated known range. In the present study, river discharges were evaluated at 13 locations within the catchment, groundwater fluctuations at 30 locations, and soil water content at 3 locations (Figure 1).

### 4.1 Evaluation of Soil Moistures with Various Land Covers

Simulated results of soil moisture data for different land covers (mire, grassland, and forest) were compared with observed values from 1 August to 31 October 2001 (Figures 5(a-c)), together with the precipitation distribution for the same period. The greater precipitation around 10 September was due to typhoons. In grassland (Figure 5(b)), where vegetation and soil structures are simpler, the simulation accurately reproduces the measured data at soil depths of 10 cm and 1.0 m. In forested areas (Figure 5(c)), the porosity changes greatly in the vertical direction owing to a more complex root distribution, which dampens the response of the observed soil moisture at 1.0 m. The simulated value excellently reproduces the observed data at 10 cm. However, it does not reproduce at 1.0 m with higher precipitation,
because porosity and hydraulic conductivity are constant in the vertical direction in this model despite the change in soil texture with depth (Yu et al., 2001). The surface soil moisture content in forest is higher than in grassland, indicating that the forest cover retains more water in soil and vegetation than grassland does, and that grassland tends to quickly lose infiltrated water to the groundwater system and evapotranspiration. Furthermore, the differences between land cover become less marked with depth. In the mire (Figure 5(a)), the simulation takes on an almost constant saturated value (≈1.0) owing to the shallow groundwater level calculated by MODFLOW. This simulated value agrees well with the observed value of approximately 0.99. This high value is characteristic of a mire because of the poor drainage at lower elevations, the almost flat surface, the peaty soil texture, and the soil’s elasticity (Kellner and Halladin, 2002).

4.2 Evaluation of Groundwater Levels

In the distribution of simulated hydraulic heads of steady-state average for the overall Kushiro River catchment, plotted with the measured values of Ohata et al. (1975) (Figure 6), hydraulic head becomes smaller as the hydraulic conductivity of the riverbed ($k_b$) becomes larger. The simulated value of $k_b = 300$ m$^2$/h agrees most closely with the measured value, and was used in the following simulations. In the vicinity of the Kushiro mire (well number: h-60

![Figure 5](image-url)

**Figure 5** Time-series of precipitation and soil moisture at three meteorological stations from 1 August to 31 October in 2001. Soil moisture at (a) mire, (b) grassland, and (c) forest, respectively. Open-circle and open-triangle show observed values at soil depths of 10cm and 1.0m, respectively. Solid-line and dashed-line show calculated values at soil depths of 10cm and 1.0m, respectively.
Figure 6 Distribution of hydraulic heads in the Kushiro River catchment. Open-circles; measured values by Ohata et al. (1975), dashed-line; simulated value without river, dash-dotted line; simulated value of $k_b=30$m$^2$/h, and lines; simulated value of $k_b=300$m$^2$/h.

to 120), groundwater almost saturates the surrounding soil, a condition that can be reproduced very well by the numerical simulation.

4.3 Evaluation of River Discharge

The simulated discharges of both the kinematic wave (dashed-line) and dynamic wave (solid-line) methods from stream network modelling are compared with observed values (open-circles) at several points in the Kushiro River catchment from May 1 to October 31 2001 (Figures 7(a, b)). In both figures, the observed discharge decreases gradually in the same way as the ground-water levels in the mountainous areas from the spring to the early summer (Ohata et al., 1975). In the late summer and fall, the discharge fluctuates greatly, depending on volume of precipitation from typhoons.

The simulation value at Kucyoro water-flow survey station in the Kucyoro River (bed gradient here is about 1/2,000), a tributary of the Kushiro River flowing into the Kushiro mire, reproduces very well the observed value in the typhoon period, and the differences between kinematic wave and dynamic wave models are smaller, showing that the dynamic wave effect is small at this point despite the lower bed gradient (Figure 7(a)). Furthermore, the base flow can be calculated correctly because the grid-based distributed model in the current research includes recharge rates, seepage, and the return flow from the ground. However, the simulation cannot reproduce the observed data during the melting period, in particular, from 1 May to 10 June, and the simulated discharge is smaller than the observed value. The integrated model in this study does not completely include the melting process of snow volume to surface flow, a necessary consideration in the future.

At the Gojikkoku water-flow survey station downstream of the main Kushiro River region (bed gradient becomes milder, about 1/1,700), the dynamic effect is important because of the Kushiro mire region’s influence (Figure 7(b)). The peak discharge comes later and
Figure 7  Temporal variation of river discharge at Kushiro River catchment from 1 May to 31 October 2001. (a) tributary of the Kushiro River flowing into Kushiro mire (Shimo-Kucyororo observation point), and (b) Kushiro River (Gojikkoku observation point). Open circles are observed values and lines are calculated values (dashed line; kinematic wave model, solid line; dynamic wave model).

milder (but with a larger volume) in the downstream regions (maximum delay is about 5 days) due to the lower discharge rate from the mire (Winter, 1988; Kellner and Halldin, 2002), which is 50%–60% of the mire’s flow rate (Ohata et al., 1975). The simulated results of kinematic wave model overestimate the observed values especially during the rainy periods. The dynamic wave model reproduces excellently the peak value of discharge in precipitation, and the backwater effect after a flood. Thus, the grid-based distributed model used in the present research is highly accurate for different topographies because it includes surface-unsaturated–saturated water processes displaying both short- and long-term components of river discharge, effective precipitation, and the kinematic and dynamic wave effects at various river slopes.

4.4 Drying Phenomena in Kushiro Mire

The NIES Integrated Catchment-based Eco-hydrology (NICE) model shows a much higher accuracy in simulating river discharge, soil moisture, and groundwater flow over the Kushiro River catchment than previous catchment models. Its use of very accurate field measurement data, MODIS data, flux tower data, and various parameters categorized by GIS and geological structure support this precision. Seasonal vegetation cover change, mechanisms of vegetation–water relations, and surface-unsaturated–saturated water processes, account for almost all of the major eco-hydrology processes of water and heat movement in a
closed catchment.

Simulation was conducted for three years (1900, 1950, and 1985) by using land cover data (Figure 8). The size of the simulated area for groundwater around the Kushiro mire is 14.0 km long by 10.0 km wide. This region is a low-elevation mire surrounded by capes and low mountains, of which about 10–20 km² floods in very wet weather. At those times, large amounts of sediment and gravel flow into the mire and where piles are formed as evaluated by the previous researches. In the simulation, all the other forcing data, soil property data, geological structure data were the same because digital data of soil texture and geological structure was not available, and because predominant changes in the underground structure had not occurred as had changes in vegetation phenology. The groundwater levels are almost identical for three years, decreasing a little near the Alnus japonica (circled area) invasion area due to the deposit of sediment load. The simulated annual-averaged soil moisture clearly indicated the drying phenomena near the circle area due to the invasion of Alnus japonica.

The eastern side of the cape is located downstream of the channelization, therefore, the coarser sediments are deposited there as the flow velocity decreases. In contrast, on the southern side of the cape, floodwater spreads out to the lowest areas, depositing the finer sediments widely. In these regions, the simulation showed that the outflow of groundwater to the river and the inflow of recharge became smaller because of the lowered groundwater levels. The simulated soil moisture takes a greater value near the Kushiro mire, especially from summer to autumn, than in the surrounded areas, with a value greater than 0.80, because the low-lying region is inundated throughout the year, which is characteristic of a mire that is covered mainly by reeds. This shows that Alnus japonica, a deciduous tree with a height of about 15 m invaded the mire, absorbed more water from the roots and transpired more to the atmosphere than original mire vegetation such as reeds, and that the soil moisture decreases dramatically in the vicinity of the Alnus japonica invasion. Furthermore, the area of lower soil moisture and lower surface temperature closely corresponds to that covered by Alnus japonica.

In this way, the simulation clearly demonstrates that this drying phenomenon is closely related to the increased influx of sediments into the area surrounding agricultural development, reclamation, and channelization of the river (Environment Agency of Japan, 1984, 1993).
Thus, the positive correlations between soil moisture and groundwater levels indicate that the inflowing sediments during flood periods form spatially distributed piles, creating a resistance to water supplied by the surroundings, and the resulting drying of the soil and the lowering of local groundwater levels, resulting in a further expansion of *Alnus japonica* in this area. The areas of *Alnus japonica* promote the deposit of sediments due to overland roots and shedding tree leaves.

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Some of the simulations were run on an NEC SX-6 supercomputer at the CGER, NIES. We also thank CGER for providing Tomakomai Flux-Tower data for validating MODIS satellite data.

**References**


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

Shin-ichi MINEO, Masashi TSUKAMURA, Masatoshi UGAJIN
NEC Corporation, HPC System Center

1. System

1) Introduction

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

The SX-6/64M8 provides a peak vector performance of 512GFLOPS (64GFLOPS per node), and has the following features:
- 512Gbytes of main memory (64GB per node)
- 8TBytes of raid disk capacity with Global File System (GFS)
- 200TBytes of Tape Library capacity (DTF-2 format)
- Internode crossbar switches (IXS) interface at 8Gbps
- 1000Base-SX interface at 1Gbps

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

![System Configuration Diagram](image)

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

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

Inheritance and expansion of the distributed shared memory architecture

The SX-6 Series inherits the vector-processor (CPU) based distributed shared memory architecture which was highly praised in the SX-4 Series, and flexibly works with all kinds of parallel processing schemes. Each shared memory type single-node system contains up to 8 CPUs, which share a large main memory of up to 64Gbytes. In a multi-node 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 operability 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 this term.

--- 2003.4.1-2003.8.19

<table>
<thead>
<tr>
<th>Job class</th>
<th>CPU limit</th>
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<th>Run limit</th>
<th>Execute nodes</th>
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<td>8</td>
<td>2nodes</td>
</tr>
<tr>
<td>vector_ps</td>
<td>192H</td>
<td>60GB</td>
<td>1</td>
<td>2nodes</td>
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<td>(vector_atm)</td>
<td></td>
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--- 2003.12.15-present (Change for the system confusion dissolution and user request)

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<td>vector_s3h</td>
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<td>(vector_atm)</td>
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</table>

2) Improvement of system use

Support towards efficient use is mainly offered by policies of the following.

i) Use and tuning of the job scheduler ERS for SX series

ERS is a product that adds the following functions to NQS: a job scheduling function for managing priorities set up for user, group and organizational units according to past usage; and a job execution control function.

Monitors the quantity of use of memory, CPUs and other system resources, and controls job execution commencement and checkpoint/restart

With NQS operation alone, there may be cases of memory swap and other overload, as well as no use of some resources. However, use of ERS enables maximum use of resources for a system within a scope that does not exceed the maximum setting, enabling job execution commencement and performance of checkpoint/restart.

Moreover, in the event of a job submission with high priority, it is possible to take over process control from a checkpoint and suspend jobs with low priority without interruption of job execution.
Fair-share Scheduling for management - by organization, group and user – of resources used in jobs and processes (Fig. 1)

With user management in Fair-share Scheduling, it is possible to have a hierarchical structure consisting of organizations, groups and users, and to establish shares in each of them (Fig. 2).

The Fair-share Scheduler gathers and accumulates the quantity of resources used by each user, performs comparisons with established shares, and determines the order of priority of a job. This function may be used for the objective of allocating usage rights of a system in line with the budgets of each research center, and the objective of preventing the monopolization of a system upon input of many jobs by one user.

With ERS operation, it is possible to have many kinds of tuning corresponding to the operation of a site. With ERS, job submission, etc. uses NQS commands, so the effect on the end user at the time of installation may be minimized.

* NQS refers to Network Queuing System, developed by Sterling Software for NASA Ames Research Center.
ii) Support of program improvement in the speed

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

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<thead>
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<td>Tuning to the routine out of which the performance has not come</td>
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<tr>
<td>agcm5 4.02</td>
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<td>flood2D</td>
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<tr>
<td>FldDynNR_Fortran</td>
<td>Use of mathematicallibrary: ASL Sauce correction</td>
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</table>

3) Future plan

The operation in a new job class was started 2003 in December. We plan to the efficiency by improving the management and fine tuning program.

3. Use of SX-6

The following figure shows supercomputer’s operation results from Apr. 2003 to Dec. 2003

![Graph showing supercomputer's operation results](image)

**Figure 3** Supercomputers’ operation results.

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## Research Programs in FY2002

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<tbody>
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<td>Climate change experiments with a high-resolution CGCM (NIES, The University of Tokyo, Kyushu University, Nagoya University, Forestry and Forest Products Research Institute, Frontier Research System for Global Change, National Institutes of Agro-Environmental Sciences)</td>
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<td>Analysis of ILAS data using a three-dimensional chemical model (NIES and CNRS)</td>
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<table>
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<th>(2) Other Researches</th>
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<td>A study on the effects of polar ozone destructions due to bromine species (NIES/The University of Tokyo)</td>
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</tr>
<tr>
<td>Stratospheric chemistry simulation (The University of Tokyo)</td>
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<tr>
<td>Application of the transport model for inverse modeling studies of the global and regional budgets of CO₂ (Frontier Research System for Global Change, NIES)</td>
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<tr>
<td>A climate simulation at the early stage of the last glacial period (Meteorological Research Institute)</td>
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<td>A study of the interaction between the atmospheric circulations in the low- and mid-latitudes with climate models (Saitama Institute of Technology)</td>
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<td>Study on projection of climate change based on new emission scenarios (Meteorological Research Institute)</td>
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<tr>
<td>Ultra high resolution modeling of the tropical air-sea interaction (Kyushu University)</td>
<td>49-54</td>
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<tr>
<td>Development of atmospheric general circulation model for terrestrial planets and related fundamental experiments on the atmospheric structures (Hokkaido University)</td>
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<tr>
<td>Atmospheric motion and air quality in East Asia (Kyoto University)</td>
<td>67-76</td>
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<tr>
<td>Numerical simulation of the polar vortex and the geostrophic vortex on the rotating sphere (Tohoku University)</td>
<td>77-84</td>
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<tr>
<td>Quasi-geostrophic vortex motions and scalar transport (The University of Electro-Communications, Tohoku University)</td>
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<tr>
<td>Development of finite-difference numerical model for studying the geophysical fluid dynamics in spherical polar coordinates (Tohoku University)</td>
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<tr>
<td>Heat and mass transfer mechanism at and below the air-water interface in the surface ocean and the effects of droplet and swell on the heat and mass transfer at the air-sea interface (Kyoto University)</td>
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<tr>
<td>International cooperative research on the management of watershed environment (NIES)</td>
<td>105-118</td>
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<tr>
<td>Ecosystem modeling in East China Sea (NIES)</td>
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<tr>
<td>Numerical analysis of aerosol distribution and regional climate variation based on regional scale meteorological/chemical transport model over the Asia domain (Kyushu University)</td>
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(As of March 2001)
Appendices

Outlines of the Research Activity Reports
(in Japanese)

Program of the 11th Supercomputer Workshop
Tsukuba, October 21, 2003
Outlines of the Research Activity Reports (in Japanese)
臭素系物質のオゾン破壊に及ぼす影響

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3. 地球フロンティア研究システム
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はじめに

フロンガスの放出量は、規制によって、1995年以降減少しており、成層圏における Cl（活性塩素）の量は、現在までにほぼピークに達したようである。一方、ハロンガスの中には、今なお放出量が増加しているものがあるのと、成層圏の Br（活性臭素）の量は、今後もうしばらくの間増加する可能性がある。国立環境研究所・成層圏オゾン層変動研究プロジェクト・オゾン層モデリング研究チームで開発を行ってきた CCSR/NIES ナッジング化学輸送モデルを使って、臭素化合物のオゾン破壊に及ぼす影響を調べた。臭素化合物の光解離の受け易さが、不均一反応と北極圏の状態を通じてどのようにオゾン破壊に影響を及ぼすかに関する解析を行った。

CCSR/NIES ナッジング化学輸送モデル

モデルには、酸素、オゾン、水酸化物、窒素化合物、炭化水素、塩素化合物、対流化合物、硫黄化合物を含む约160種類の気相化学反応と13種類の不均一反応が導入されている。水平分解能はT21（5.6×5.6°）、鉛直方向の分解能は2～3 kmである。極域の光化学に重要な、物質の光解離係数の計算には、大気の球面形状を考慮している。図1に、1997年と2000年のキルナ（スウェーデン）、札幌、つくばにおけるオゾン全量観測値と、化学輸送モデルによる計算値の時系列を示す。モデルは、極圏内部の不均一反応過程によるオゾンの破壊と、極圏の移動・変形によるこれらの地点の急激なオゾン変動をよく再現している。
図1 キルナ（スウェーデン）、札幌、つくば上空のオゾン全量の観測値（EP-TOMS、ピンク）と計算値（青）。縦軸はオゾン全量（ドプゾンユニット）、横軸は1月1日からの日数を表す。

結果
(1)臭素化合物に関係する不均一反応の役割（図2）
不均一反応を導入した場合と導入しなかった場合との計算結果の比較により、不均一反応が介在することによって、オゾン破壊が始まる前の冬の期間に、成層圏下部ではBrONO₂に代わってBrClの蓄積が進むことがわかった。BrClはBrONO₂よりも弱い太陽光で光解離するため、春先の太陽光の増加に敏鷹に応答してBrOの急増を起こし、BrO-ClOオゾン破壊サイクルを促進する。

(2)極渦の状態と臭素化合物によるオゾン破壊との関係
北極渦が安定に存在し、5月上旬まで持続した1997年は、高緯度の下部成層圏に高濃度のClOとBrOの両方が存在して、ClO-BrOオゾン破壊サイクルが有効に働いた。

(3)臭素化合物によるCl₂O₅サイクルの強化
臭素化合物を導入した場合としなかった場合との計算結果の比較により、不均一反応によるBrClの増加によって、塩素化合物も活性化され、Cl₂O₅オゾン破壊サイクルが強化されることはわかった。

図2 (左)不均一反応過程を化学輸送モデル導入した場合の臭素化合物の濃度（体積混合比）の季節変化。1997年、高度45OK、80°N、経度平均の値。縦軸の単位は、ppmv、横軸は、1月1日からの日数を表す。(右)不均一反応過程を導入しなかった場合。
1. はじめに

海洋の全球規模塩循環は気候の平均的状態や長期変動をコントロールする主なプロセスのひとつである。それ中でもグリーンランド海とラプラドル海における深層海水形成に端を発する大西洋オフショルト循環（AMOC）は最もも重要なものであり、その強度や安定性は北半球高緯度海洋の淡水・塩分収支に大きく依存する。

この研究においてはAMOCに対する北半球高緯度淡水収支の重要性に関して、ふたつの側面に着目する。ひとつは広範で用いられている海面淡水フラックスの数値データセットの違いが海洋大循環モデル（OGCM）で再現されるAMOCに対して及ぼす影響であり、もうひとつは比較的低塩分の太平洋水の北極海中の通過道がAMOCに対して及ぼす影響である。

2. モデル

ここで用いるOGCMはCCSR Ocean Component Model (COCO) バージョン3である。これは自由海面を表現したリアルタイムモデルで、粒子間交換の水深方向にのみモデルが適用されている。このモデルは多くの最新数値計算アルゴリズムや物理パラメタリゼーションが含まれている。淡水モデルは2カテゴリー表現を採用しており、各水平格子において塩度と平均厚さが予報される。海水モデルの熱力学は最も簡単な0層であり、力学は粘弾性レオロジーに基づいている。

モデルを駆動するすべてのフラックス（熱・淡水・運動量）は、特に述べる場合を除き、ヨーロッパ中西部天候予報センター（ECMWF）再解析データに基づく数値値から導かれる。このデータセットは海洋モデル比較検討プロジェクト（OMIP）において採用されているものである。伝統的な海面塩分の気候値への緩和は、この研究では行わない。これによりモデルの塩分場にはバイアスが生じてしまうが、淡水収支の深層海水形成や熱循環における役割を誠実に論じるためには避け得ない選択である。

3. 結果

3.1 淡水フラックスデータセットの違いのAMOCに対する影響

海面淡水フラックスの気候値データセットには、特に河川流出と熱帯降水中の大きな不確実性があり、モデルでシミュレートされるAMOCはそれらの違いに対してとても敏感度が高いため、図1においてはOMIPデータセットで駆動して得られるAMOCと淡水フラックスを別のデータセットに差し替えて得られるものと比較している。このもう一方の淡水フラックスでは、蒸発と降水に関してはNCEP/NCAR再解析を、河川流出に関してはPerry et al. (1996)が編集した直接観測値を用いている。

河川流出データを差し替えるだけでも図1に見られるのとほぼ同じ程度のAMOCの違いが得られる。多数の数値実験を通じ、モデルでシミュレートされたAMOCは北半球極域海洋に対する河川流出データの違いに対して特に敏感度が高いことがわかった。北半球高緯度の河川流出は全体の13から14％を占めるのみであるが、それがモデルでシミュレートされたAMOCの違いの40％以上を説明する。蒸発・降水フラックスを差し替えるAMOCには顕著な違いは現れない。しかし、蒸発と降水の差を、その背景にある物理を考慮に入れてさらに分解すると、そのそれぞれAMOCに対して大きなインパクトを持つ。
3.2 太平洋水の北極海中の通り道のAMOCに対する影響

北極海は主として河川流出とベーリング海峡通過流から淡水を受け取り、それらは最終的にカナダ多島海を抜けてラプラドル海へ、あるいはフランク海峡を抜けてグリーンランド海へ排出される。したがってこの淡水の通り道はAMOCに影響を及ぼす。ここではカナダ多島海の通り道を開けた実験（OPEN）と閉じた実験（CLOSED）の比較を行う。

AMOCの体積輸送流線数の赤道における極大値は、OPENでは12.4 Svであるのに対し、CLOSEDでは8.7 Svである。CLOSEDに比べてOPENではラプラドル海とグリーンランド海両方の深層水形成が活発化されている（図2）。グリーンランド海での活発化の理由は単純である。OPENではベーリング海峡を抜けてきた塩分の低い太平洋水の多くがカナダ多島海を抜けるのに対し、CLOSEDではそれはすべてフランク海峡を抜けなければならない。したがってグリーンランド海の塩分はOPENの方が高くなり、深層水が形成され活発化される。一方、OPENでは低塩分の太平洋水がカナダ多島海を抜けるのであるから、ラプラドル海での深層水形成は弱まることが予想されるが、実際にはそのような状況はなく、モデルにおいては太平洋水はラプラドル海の深層水形成領域を取り巻く低気圧循環に取り込まれ、深層水形成領域に直接影響を及ぼさない。その代わりに、グリーンランド海で塩分が高まったことの影響が東グリーンランド海流および西グリーンランド海流によってラプラドル海の深層水形成領域に及び、深層水形成が活発化されている。

図1 大西洋における年平均東西積分体積輸送流線数。OMPデータセットで駆動された場合（左）とNCEPデータセットで駆動された場合（右）。等価線間隔は1 Sv。

図2 鉛直座標をσtとして描いた大西洋における年平均東西積分体積輸送流線関数。 (左) OPEN, (右) CLOSED。等価線間隔は1 Sv。
炭素循環インパースモデルに関する並列大気輸送モデルの開発及び適用

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1. 目的

人為起源のCO₂濃度の観測データは、陸域生態系と海洋の両方の炭素循環関連の情報を多く含んでおり、炭素循環過程の解析にとっては有益な情報となる。実際の観測データを大気輸送モデルに組み込んで、大気表面CO₂フラックスの季節・年々変動と時間分布を解析した報告数例がある。しかし、インパース大気輸送解析の精度は、フォワード輸送シミュレーションや表面フラックスのインパースモデルの空間的・時間的分解能に依存している。本研究では、クラスター・ベクトルスーパーコンピュータ上に最適化された並列大気輸送モデルを用いて、空間的・時間的に高分解能のインパースモデル解析手法の開発を目指した。

2. 概要

インパースモデルによる炭素循環の解明において、CO₂フラックス分布情報は、大気CO₂観測と表面CO₂交換の空間・時間分布(Enting et al., 1995, Bousquet, 2000)に関する既存情報により補正することができる。インパースモデルにおける補正の精度は、輸送モデルの精度・解析方法や発生源の空間分解能や解析する時間分解能に依存する。後者の影響はKaminski et al. (2001)によって集約エラーとして記述されており、発生源の領域の分割数はコンピュータの性能が許す限り、大きい方が好ましいとされる。本研究において用いる大気輸送モデル(Mksyutov and Inoue, 2000)は、月・年単位のトレーサーバランス関数のシミュレーションに使用されてきた。それをMPIインタフェース機能を持つマルチプロセッサスーパーコンピュータを用いて並列計算を実現するために、輸送モデルに幾つかの修正を施した。ここで、NIESのスーパーコンピュータシステムは、モデル開発と超大型スーパーコンピュータシステムである地球シミュレータでモデル計算が実行できるようにするための事前調整・準備に使用した。

輸送モデルはマルチプロセッサで効率的に流れれるよう再設計し、NECスーパーコンピュータの72プロセッサを用いることにした。これにより、トレーサーの異なるグループ輸送モデルが各プロセッサを基に独立に実行され、良いスケーラビリティとパフォーマンスを提供できるようになった。また、スケーラビリティ達成のために2つのプロセッサを試した。その1つは、各プロセッサ要素が風の場を読み込んで処理をするものである。他は、1つのプロセッサ要素が読み込んで処理し、メッセージバッキングメカニズムを経由して、他のプロセッサ要素がその前処理されたデータを受け取るものである。双方のアプローチともに、計算速度が向上した。しかし、このバフォーマンスが可能たのは、高速・大容量並列コンピュータのみであった。同様のNEC SXシステムを用いて行われた詳細なインパースモデル研究は、他に一例、ハンブルクのマックスプランク気象研究所におけるRodenbeck et al. (2003)によるものがあるのみである。

本研究では、最近20年間の炭素循環を解析することを目標として、Rayner et al. (1999)の報告に類似したトランスコム3の国際インパースモデル解析手法を用いて、NCEP再解析データの風の年々変動データを使用したフォワードシミュレーションによって、64のインパースモデル領域(42陸域、22海域)をそれぞれに対する月平均フラックスを解析した。輸送モデルは鉛直変動に対し15σの精度、水平方向には2.5×2.5の分解能をもたせた。インパースモデル解析にはNCEPの圧力レベルデータを使用したが、100mbar以上の鉛直速度のデータ不足分に関しては、等エントロピー等温度等速度ジェストリで解析されたものを補った。
陸域・大気及び海洋・大気間フラックスの季節及び年々変動は、実際にの大気 CO₂観測データや人為的影響のパターン図、陸域・海洋フラックスデータを使用して、月間フラックス補正に Bayesian インバースモデルを適用して得た。

本研究の他の目的は、炭素循環インバースモデル解析に将来の衛星観測データが適用できるか否かの検討である。リモートセンシング観測によって得られる大量のデータに対して限定された発生源の領域を使用したインバースモデルでの CO₂発生源と吸収源の推定において、地上ベース観測と衛星観測との効率性を比較した。その結果、地上ベース観測の方がより正確なものであることが分かった（Patra et al., 2003）。

すなわち、衛星観測データを利用した解析のためのインバースモデル領域の小領域化、領域細分化することが基本となる。15 度間隔の緯度－経度領域を 10 グリッド、トータル 432 領域に対してインバースモデルを使用して、フラックスの不確実さを検討した。年間平均モデルにおいて、432 領域の年間データ（パルス）を解析することによって、インバースモデルの不確実さが低減した。その詳細は Maksyutov et al. (2003) に報告されている。

3. 成果

64 領域に分割した陸域及び大気 CO₂フラックスにおける季節・年々変動の解析は、年間、あるいは10年間の時間スケールでの大気 CO₂フラックス変動と気候変動の関係を解明するのに役立つことが分かった。本研究での解析結果では、陸域（主に熱帯域）及び海洋フラックスに対するインバースモデルフラックスが南方振動の指標（SOI）に相関があることを示した。

一般的には、エルニーニョ発生期中はより大きな海洋吸収（CO₂の真の吸収源）及び陸域からの放出（CO₂の真の発生源）があると考えられた。真の海洋の吸収源には、主に赤道付近太平洋洋域における低気圧の小ささが関係する。陸域の吸収源は大気の乾燥状態が引き金となっており、例えば、1997～1998 年のインドネシア森林火災のように、エルニーニョ発生期中には森林火災が促進された（図 1 参照）。

衛星データ利用の研究評価において、インバースモデルによるフラックスの空間分布の不確実性は、1）反射太陽光の高度分布観測、2）700mbbar 以上での任意の云域観測、3）海洋上における晴天時の反射太陽光の高度分布観測の 3 セットの空間観測から評価した。ケース 1）の高度分布観測から得られた情報量は最も多く、フラックスの不確実さの低減を示した図を、最近の出版物（Maksyutov et al., 2003）に詳細に記載した。
気候変化に対する熱塩循環の長期応答
—理論的考察とCGCM実験—

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1. はじめに
結合モデルを用いた各種気候変化実験における大西洋熱塩循環の振舞を解釈するために、ボックスモデルを用いた理論的考察を行った。Rahmstorf の4-ボックスモデルにおいて、淡水及び熱に関する（外的）強制力を独立あるいは同時に変化させた場合の平衡解の応答を数学的に調べた。また、現在の熱塩循環がどのレジームにあるのかを調べるための結合モデル実験を行った。

2. 理論的考察
図1にRahmstorf (1996) により導入された4-ボックスモデルの概要を示す。ここでボックス1は大西洋低緯度を、またボックス2は北大西洋の高緯度を表していると考える。このモデルの$q > 0$である平衡解は以下の関係式を満たす。

\[ q^2 = -q k \Delta T + k \beta T_0 F_a \]

(1)

この関係を$F_a - q$ 垂直位相図が図2に示したのが図2に示したレジームダイアグラムである。これらの平衡解は 1) 熱駆動レジーム、2) 熱塩駆動レジーム、3) 塩分駆動レジームのいずれかのレジームに属し、レジーム毎に外的強制の変化($[\Delta T^*], [F_a]$) に対する応答は異なる。それらを理論的に解析した結果が表1に示されている。

![Rahmstorf の4-box モデル](image1)

![Atlantic THC regime diagram](image2)

図2 4-box モデルのレジームダイアグラム
3. 結合モデル実験

表1によれば、熱塩循環のどのレジームにあるかは、淡水による外的強制力のみを変化させた場合の、南北両ボックス間の塩分濃度差の変化傾向に特によく表れる。そこで、そのような強制力の変化を実現できるような結合モデル実験を企画し実施した。表2にその結果を示す。ここで、ReWt実験は、淡水による強制力のみを増加させたもの、またRtWc実験は熱的な強制力のみを減少させたもの、tran実験は両者を同時に変化させたものである。

<table>
<thead>
<tr>
<th>forcing</th>
<th>thermal ($F_0 &lt; 0$)</th>
<th>thermohaline ($F_0 &gt; 0$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$F_0$</td>
<td>$\Delta T^*$</td>
<td>$q$</td>
</tr>
<tr>
<td>/</td>
<td>/</td>
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</tr>
</tbody>
</table>

表2 結合モデル実験の結果

<table>
<thead>
<tr>
<th>run</th>
<th>$q$</th>
<th>$S_2$</th>
<th>$T_2$</th>
<th>(forcing)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ReWt</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>$F_0$</td>
</tr>
<tr>
<td>RtWc</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>$\Delta T^*$</td>
</tr>
<tr>
<td>tran</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>$F_0$ and $\Delta T^*$</td>
</tr>
</tbody>
</table>

4. 結論

表1に示した、理論的な結果と表2に示した結合モデル実験の結果を見比べると、実験に用いたモデルの熱塩循環は熱塩駆動レジームにあると判断される。他の数多くのモデルの実験結果や観測における大西洋の淡水収支解析結果などとも照らし合わせると、現実の大西洋熱塩循環は熱塩駆動レジームにある可能性が高い。この結論に基づくと、温暖化実験の長期ランにおける熱塩循環の振舞(Stouffer and Manabe 2003)や、古海洋データから判断される氷期における熱塩循環の振舞を整合的に説明することができる。
雲解像モデルによる放射対流平衡

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1. 研究目的
放射対流平衡は大气構造を理解するための基本概念であるだけでなく、温暖化問題における気候変動の観察に必要不可欠である。近年では、積雲対流と放射過程の直接的な計算が可能な積雲解像モデルを用いた放射対流平衡計算が行われるようになってきており、対流の循環構造と大気の平衡状態との関係を力学的に考察する必要性が増してきている。積雲解像モデルによる放射対流平衡計算は、次のような役割をもっている。
1) 放射と雲の相互作用の直接的な計算による気候変動評価の精密化。
2) 積雲対流過程の自然に計算できるため、積雲対流層の微細構造の解明。
3) 低分解能モデルの結果との比較による積雲パラメタモデルの検討。
4) 積雲解像モデルにおける雲物理、放射、乱流などの物理過程の特性の評価。
5) 積雲解像モデルの他のモデルとの比較評価。

一方で、多くの積雲解像モデルによる放射対流平衡計算事例があるものの、実験設定の違いのため意味ある比較が困難になっている。今後、温暖化予測に重要な雲対流相互作用を正しく評価するためには、モデル間の差異によらないrobustな性質について理解を深めていく必要がある。本研究では、積雲解像モデルによる放射対流平衡計算の標準実験を提案し、その問題点を検討したい。

2. 標準実験設定と論点
実験設定はTompkinsとCraig (1998)を参考として次のような条件を標準実験の条件設定とする。

- 次元数: 3次元。
- 計算領域: 100km × 100km, 高さ: 20km。
- 格子間隔: Δx=Δy=2km。
- 鉛直層: 30層, 下層ほど細かい。
- 境界条件: 周期条件。
- 下端境界条件: 一様海面条件, 海面温度300K。
- コリオリパラメータなし, f=0。
- 放射: −2K/day の一定放射。
- 地表面フラックス: 最小風速4m/s。
- 大規模Forcingなし。

この条件を基準にさまざまなパラメータを変えて平衡状態を求めた。結果を、まず質量重み付き平均温度, 加温水量のダイアルグラム上で比較する。

3. 実験結果
図1には標準実験と雲物理過程を変えた実験の平均温度, 可降水量を示す。CTLが標準実験でKessler型の暖かい雨を用いた。雨の成長にBerryの方法を用いた結果と, RobeとEmanuel (1996), Grabowsky (2003)にしたがって雲水からの雨への変換過程を変更した結果をそれぞれRE96, G03として示す。図2にはCTLとRE96の相対湿度の高さ分布を示す。CTLでは上層が飽和状態であるのに対し, RE96は飽和状態が解消されている。この結果のほうがより現実に近いと考えられる。

このことは, 雲物理過程の變化が熱帯の平衡状態に重要な役割を果たしていることを意味する。
一連の実験については, 熱帯条件に比, 平均温度が低く, 可降水量が小さい。これにより, 地表面温度と海面温度に大きなギャップが生じることに由来する。地表面パルク係数を大きな値に設定するか, 最小速度を大きくすることによって現実に近づけることができる。
図1 質量重み付き平均温度と可降水量の依存性。
CTL：標準実験，雲物理過程を変えた実験結果。

図2 相対湿度の鉛直分布。左：標準実験 CTL，右：RE96。
新排出シナリオに基づく新しい気候変動シナリオの推計に関する研究

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1. 背景と目的
気候変動に関する政府間パネル(IPCC)の次期評価報告書(AR4)作成に向けて、国立環境研究所(NIES)のアジア太平洋地域統合モデル(AIM)グループやシナリオ作成事務局の作業グループ(TGCCIA)、第一作業部会の技術支援事務局(TSU)の要請にもかかわらずシナリオ実験を行うことが本研究の目的である。予測シナリオは本来排出シナリオの形で提供されているが、大気中にはば一様に分布している温室効果気体については濃度シナリオでも提供されている。地球温暖化予測に使われている現在の気候モデルには化学輸送モデルは組み込まれていないので、実験を行うために大気中の濃度分布を与える必要がある。このため、気象研究所で開発された化学輸送モデルを用いて硫酸アセロゾルの濃度分布を作成し、そこで求めた次の元のアセロゾル分布を用いた温暖化予測実験を開始した。

2. アセロゾル分布の計算
大気中の硫酸アセロゾル濃度分布の計算には、気象研究所で開発された大気循環・対流圈アセロゾル化学輸送結合モデル(MASINGAR: Tanaka et al., 2003)を用いた。アセロゾルの移流計算には三次元のセミラグランジュ法を用い、発生、乱流拡散、積雲対流による輸送、重力沈降、乾性沈着、湿性沈着を考慮してアセロゾル三次元分布を計算した。計算にあたっては、大気・海洋結合モデルの基準実験（放射強制力長期積分値を基にした長期積分値）の6時間おきに出力された気象場による緩和（ナッシュ）を行った。

二酸化硫黄の排出量は年平均値を与えているが、計算された硫酸アセロゾル分布には大きな季節変動が見られた。図1に二酸化硫黄の排出量、硫酸アセロゾル量、直接効果による放射強制力を示す。計算された人為起源の硫酸アセロゾルの総量は1990年で0.79TgS、全球平均の直接効果による放射強制力は-0.54W/m²となっている。

図1 a) 二酸化硫黄の排出量、b) 硫酸アセロゾル量、c) 直接効果による放射強制力の時間変化。a), b) については6つの全てのSRESシナリオ(A1B, A1FI, A1T, A2, B1, B2)、c) についてはA1B, A2, B2のSRESシナリオにおける計算結果。
3. 三次元のエアロゾル分布を用いた温暖化予測実験

温暖化予測実験には気象研究所で開発された大気・海洋結合モデル(MRI-CGCM2: Yukimoto et al, 2002)を用いる。このモデルの解像度は、大気が水平T42鉛直30層で0.4hPaを大気の上端とするハイブリッドα-p座標、海洋は東西2.5°、南北2°、鉛直23層で熱帯域は南北0.5°となっている。放射過程については、二酸化炭素(CO$_2$)、水蒸気(H$_2$O)、オゾン(O$_3$)、メタン(CH$_4$)、亜酸化窒素(N$_2$O)は温室効果気体として直接扱っているほか、硫酸エアロゾルによる放射強制力については直接効果のみを扱っている。

実験に際し、温室効果気体については、CO$_2$とCH$_4$、N$_2$OはSRESの濃度の値をそのまま与え、フロン等の他の温室効果気体については同じ放射強制力を持つCO$_2$の量に換算して加えている。硫酸エアロゾルの分布についてはCTMで求めた分布を時間内挿して与えている。大気および海洋の初期値については20世紀の気候再現実験における1990年のリスタートデータを使用している。

図2に全球平均の地上気温と年平均降水量の現在(1961年から1990年の平均)からの変化を示す。A1B、A2、B2シナリオにおける今世紀末の温度上昇はそれぞれ約2.4℃、2.0℃、2.7℃となっている。硫酸エアロゾルの分布はシナリオ間に関なり相違があるが、温暖化したとき地上気温の変化は非常に似た地理分布をしている。

図2 MRI-CGCM2を用いた歴史実験とSRESシナリオをもちいた21世紀気候変化予測実験の結果。a) 全球年平均地上平均気温とb) 全球年平均降水量の1961-1990年平均からの偏差。
熱帯大気海洋相互作用の超高解像度モデリング：大気降水解像モデルの自然変動

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1. 背景と目的
エルニーニョに代表される熱帯大気海洋相互作用は、大小さまざまなスケールで生じている。本研究では、この相互作用の素過程を広い領域でかつ分解能が高い数値モデルで直接計算することを目指している。本年度は、モデル中で自発的に生じるゆっくりした変動についてしらべた。

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

3. 成果
図1 はモデル中の色々な平均量の時間変化を示す。全ての変数に15日程度に時間スケールでの揺らぎが見られる。色々な変数の系譜だった変動は、この揺らぎが雲活動・大規模な風の変動・海からのエネルギー供給の間の相互作用から起こることを示唆する。

図1 領域平均量の時間変化 (a) 高度3700 m の温度偏差, (b) 高度50 m の水蒸気混合比, (c) 全運動エネルギー, (d) 蒸発量, および, (e) 降水量
図2 はモデル中で時間を追って雨がどこに降るかを示している。雲は全般に東に進む大規模波動の中で多く生じているが、降水量の極大期には、雲活動は2,000–3,000 km のスケールを持ち、東西に伝播する少数の雲システムに集約される。現実の熱帯大気では、降水活動や風の場などが数十日のスケールで変動する現象が「季節内変動」として知られている。通常、季節内変動については、いくつかの変数のパターンが東進することが強調されているが、実は、熱帯全体の平均温度なども揺らいでいる。本研究の結果は、こうした現象についての有益な情報を与える可能性がある。

図2 降水量分布の時間発展。縦軸は時間、横軸はモデルの水平軸。東西に伝播する幅2,000–3,000 km の活発な雲システムを赤・緑の楕円で記す。
地球型惑星大気大循環モデルの開発と基礎的実験

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1. 目的
地球型惑星に適応可能な大気大循環モデル(GCM)を構築することを目標として、惑星大気モデルのプログラム構造の設計と我々が提案するコード書法に基づいたモデルの実装実験に着手した。それに並行して、地球および惑星大気を念頭にいただいた大気循環の基礎的実験を行った。

2. 惑星大気大循環モデルの開発に向けての実装実験
地球型惑星大気大循環モデルの開発に関しては、ソースの可読性が高く、計算設定およびモデル内で使用される物理過程の可変性が高いモデルのあるべき姿を模索した。可読性と変動の高い数値モデルの形を模索するべく、我々のグループでは階層的地球流体力学スペクトルモデル群(SPMODEL; GFD-Dennou Club, 2001)の開発と整備を継続して行っている。SPMODELでは、スペクトル解読と微分演算を行う関数を用意することにより、変数配列の格子点情報の観察している。今年度は、惑星大気モデルの力学解析の縮小システムである球面浅水方程式スペクトルモデルの実装実験を行った。時間方向の離散化には leap frog スキームを用い、計算モード除去のため Asselin の時間フィルターを組み合わせている。このとき、例えば流速の時間積分部分はソースコード内で以下のよう記述される。

```fortran
do it=1, nt
  w_ZetaA = w_ZetaB + 2 * delt * 
  (- w_Div_xy_xy((xy_Coli + xy_w(w_Zeta)) &
    * xy_w(w_U)) &
    (xy_Coli + xy_w(w_Zeta)) &
    * xy_w(w_V)) / R0)
end do
```

ここで w_Div_xy_xy( ), xy_w( ) はそれぞれSPMODELで用意された2次元流およびスペクトル逆変換を計算するための配列を返す関数である。SPMODELを用いることにより、数値モデルのソースコードは数式との対応が容易化した可読性の高いものとすることが可能となった。

3. 気候の太陽定数依存性：雪玉地球実験
地球型惑星大気を念頭におわら基礎実験としては、表面温度に依存してアルベドの値が変わる水惑星において太陽定数の値を変化させるパラメータスタディをおこなった。従来、Budyko (1969) やSellers (1969) のエネルギー流モデルで議論されていた気候年次トピックのGCM版を完成させることができた。その後の結果を以下に示す。

図1は、様々な太陽定数を与えてGCMの計算を行った結果である。GCMでは22度よりも低緯度に水境界が存在する安定した平衡状態が得られなかった。この結果は、3次元系でも大気球安定解を求めるのは困難である。一方、氷面積が減少した場合には、GCMでは統計的平衡状態が得られた。このことは、EBMと同様に小極冠不安定が起きることを意味している。しかし、GCMの計算結果では、大気界がしっかり存在するわけではない、太陽定数の値が増大するとともに、氷境界の振動の振幅は大きくなってしまい、この結果は以下のような可能性があることを示唆している。3次元系では安定振動解と時間発展問題では得られない
不安定平衡解が存在しており、EBMにおける小極冠不安定の発生のGCMにおける対応物はHopf 分岐である。

図1 GCMで得られた海水の境界線度と太陽定数の関係：○は全球凍結平衡解を、●は全球凍結していない平衡解を、×は暴走温室状態を示す。Fの文字が付いたものは$S=1000\;\text{W/m}^2$で得られた全球凍結状態を初期値とするものである。Rの文字が付いたものは$S=1600\;\text{W/m}^2$で得られた暴走温室状態を初期値とするものである。Pの文字が付いたものは$S=1300\;\text{W/m}^2$で得られた部分凍結状態から徐々に太陽定数を減少させていて得られた解たちはある。文字がついていない記号は280Kの等温状態を初期値とする結果を示す。

4. 水惑星暖水域実験：赤道上に2つの暖水域を置いた場合

地球大気を念頭に置いた基礎的実験としては、東西一様南北対称な海面水温を持つ水惑星の赤道上に2つの暖水域を置いた実験をおこなった。図2には、赤道上に2つの暖水域を置いた場合の降水量の平面分布を示す。暖水域の東側ではケルビン波的な応答に伴う赤道への水蒸気収束及び赤道上の降水増加。暖水域の西側ではロスビー波的な応答に伴う赤道から亜熱帯への水蒸気発散及び赤道上の降水減少が起こっている。暖水域東側の降水正偏差には、表面気圧における負偏差・対流圏中層の温度の正偏差が伴っており（図は示さない）、ケルビン波的な特徴を示している。この結果は、暖水域が2つの場合でもToyoda et al (2003) と同様に、熱帯域の降水および循環場の応答は赤道波の力学によって記述可能であることを示している。

図2 2つの暖水域を置いた場合の降水量分布(W/m²)。暖水域無しの標準実験からの偏差を示す。実験の最後の700日間の平均を示す。
大規模山岳風下域に発生する渦の数値実験

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1. 背景と目的

回転層大気、すなわち低ロスピー数（Ro<<1）の環境場に、大気の厚さに比べて山の高さが無視できないようなスケールの大きい山岳が存在することにより、山岳まれやその風下側にはさまざまな特徴や現象があらわれる。Happert and Bryan(1976)は回転層流体において、非線型理論および数値計算により山岳上で山岳と同規模の低気圧性渦が発生し風下へ移流する様子が再現された。似たような現象は Boyer and Zhang(1990)による実験実験においても、周期的な流速を与えることによって断続的に低気圧性渦の生成・移流が起こることが示されている。そこで本研究では三次元非静力学数値モデルを用い、大規模山岳を置いた回転層大気中において、さまざまな一般風及び大気安定度、およびコリオリ f を与え、風下側に発生する低気圧性渦の生成・移流の再現を行い、パラメータ同定を試みた。

2. 概要

用いた数値モデルは三次元ブラミネト方程式系非静力学モデルである。ドメインは6000 km x6000km x25.0kmであり、f 平面のコリオリ力を入れ、山は高さ 4000m のベル型の山を置いた。一様な一般風を設定しつつさまざまな風速、さまざまな大気安定度、さらには山岳のパラメータを何段階かに分け計算を行った。高度約10km までの環境中では、ロスピー数は0.1のオーダー、フルード数は0.1~1.0のオーダーである。

山岳を含む領域を Hunt et al(2001)に従い2つに分け、渦生成について調べていくことにする。この領域は、山岳を越えることができる流体領域である top layer、山岳を越えることができない流体領域である middle layer の2つである。まず top layer では低気圧性渦の生成は主に大気柱が伸びることによる渦度生成であり、middle layer での渦生成は主に山岳壁面の摩擦や傾斜の生成である。そこでまずは top layer における低気圧性渦の渦度と Fr 数や Ro 数など無次元パラメータの関連を調べた。次に middle layer において渦生成の有無を検出し、渦の構造を調べた。

図1 top layerにおける低気圧性渦の生成の様子。

図2 各計算における無次元渦度をFr数、山岳アスペクト比などの無次元数を使って表現したもの。Fr数（x軸）で表現したダイアグラム。
3. 成果

図1はtop layerにおける低気圧性の様子を示したものであり、このためを各計算において無次元化し(\( \text{Fr} = Fr^{1/2} \times Re^{1/2} \times hm / L^{3/2} \times \mu^{*} \))プロットしたものが図2である。Fr<0.3ではほぼFr数に比例するものに対し、0.3<Fr<3.0ではほぼ一定の値をとっていることがわかる。ここで重要なことは、コリオリ力が含まれていないことである。大気のストレッチ効果により渦生成が行われているが、それはコリオリパラメータには依存しないことが計算により明らかになった。この渦度についての解釈をするため、Huppert and Bryan (1976)を参考に準地衡風に基づくパラメータを同定を行った。その結果、Fr<0.3における数値計算結果については大体の一致を示した。

次にmiddle layerにおける渦構造の特徴を詳しくみるため、まず各計算において渦の検出をおこない、Fr数とRo数のダイアグラムにしたものが図3である。○は渦生成がありしかも長寿命のものであり、△は渦の生成が確認されなかったものである。一般に高Ro数、低Fr数であればwake領域が形成されるが、この結果ではRo数が十分に小さいとwake領域が形成されないことがわかる。また、○×の分布はおおよそ連続的である。この結果を理論的に解釈するために傾圧不安定の理論を適用した。嶋能風下限において妥当な要素のスケールを仮定し、発達条件を計算した。この結果を図3の太線で示した。太線の左は傾圧に安定していない領域であり、左下は傾圧安定の領域である。図を見ると、渦形成の有無と傾圧安定の曲線はほぼ対応している。傾圧に安定していない領域の渦構造は、渦度の軸が傾斜に対して傾いており、定常であった。しかしながら傾圧安定波であるというよりはむしろ慣性山岳重力波による定常な傾斜であるような構造であった。一方傾圧的に安定な領域での渦構造はほぼ傾斜に立った軸を持っており、その軸は山岳壁面近くから成される様子がわかった。渦度変数の解析により、この渦生成は摩擦項が主原因であることが判明した。この計算条件では、top layerにおける lee waveの傾斜運動はmiddle layer内部にあまり影響を及ぼしていないことが示唆される。この結果から渦構造の概念図を図4に示した。図1は傾圧的に安定な領域での渦構造を示したものである。Middle layerでの lee vortexとtop layerでの lee waveは独自の軸を持ちており、lee vortexが存在し得る状況となっている。一方右図はlee waveとlee vortexが合せて3次元的な構造をしている図である。

これらの結果から、top layerでの渦形成はFr数や山岳アスペクト比、無次元山岳高さにより、Ro数にはならないこと、middle layerではFr数とRo数に依存して渦形成の有無が図4のような構造をもつ決定されることが明らかになった。

図3 各計算での渦の有無をFr数（横軸）、Ro数（縦軸）の位置で表示したもの。○は渦の寿命4.0でかつその間に水平距離L以上移動したもの、△は寿命4.0で移動距離がL未満のもの。△は寿命1.0のもの。

図4 2次元的な構造を持つ渦と3次元的な構造を持つ渦の概念図。
Counter-Rotating 準地衡風梢円体渦対

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1. 序論
大気や海洋といった地球流体中には大小様々な渦が出現し、その相互作用が乱流力学を支配する。近年、これらの渦を梢円体で近似した乱流渦モデル（いわゆる梢円体渦モデル：EMM）が開発された。個々の渦は一様ポテンシャル渦度をもつ梢円体によって近似されている。ここでは、異符号の2体の渦対（双極渦）の運動を調べる。安定な渦対は長時間平行移動するので、地球流体中における‘長距離スカラー輸送’で重要な役割を担う。

2. 梢円体渦モデル（EMM）
一様ポテンシャル渦度 q1=-q2=1 の二つの梢円体渦の運動を考える。ここで、渦度の中心は (X1,2, Y1,2, Z1,2) にある。ポテンシャル渦度は両方の梢円体渦内部では一様であり、主軸長は α1,2, β1,2, γ1,2 によって記述される。これら梢円体渦の原点は、オイラー角 φ1,2, θ1,2, ψ1,2 で表される。時間発展はEMM方程式によって支配される。渦のアスペクト比は α1,2/γ1,2=β1,2/γ1,2=0.3162 とし、偏長梢円体渦は初めに鉛直に立っている。渦高さは1とし、渦はa水平方向に、h鉛直方向に離されている。

図1 EMMによる計算結果
図2 角度 θ (θ1=θ2)
(a, h) = (0.8, 0.4), 領域 (1)

図1では三つのパターンが記述されている。ケース（1）はyの正方向へ安定に平行移動する場合、ケース（2）は大きな偏角運動をしながらyの負方向へ平行移動する場合、ケース（3）は両方の渦が傾いて倒れる場合である。ケース（1）の領域における運動は周期 46.2 と 6.18 で倍周期である（図2：傾斜角 α=θ2）。振幅が小さくより早い振動が、主となる偏角運動に重ねられる。二次的な振動は、最も長い主軸（γ1,2）に垂直な平面内におけるアスペクト比（α1,2/γ1,2）との時間変化を結びつける。

ケース（2）では、図3（a=0.6, h=0.95）のように傾斜角と方位角は両方とも非周期的に変化する。t=160 になるまで傾斜角は増加し、主軸γ1が最も大きくなる。t>160 になると、γ1は減少し、α1が1のオーダーを保ち、β1が0へと減少する。つまり梢円体渦は
バンケーキのような形になる。ケース (3) では、両方の渦が無限に引き伸ばされる。この特異なふるまいのため、EMM計算はケース (3) で計算を止めてしまう。これはEMMの深刻な欠点である。ケース (2) と (3) の間の境界線はおよそ h/a=1.41 である。方位角φ1=-φ2 は明らかに臨界値へと向かう（図4：(a, h) = (0.8, 0.4)）。

3. CASL 法

楕円体の変形と散逸の影響を調べるために、CASLに基づいて、EMMの予測する3ケースに対応する数値計算を行った。

ケース (1) では、時間発展させる二つの渦は並進し、安定的に運動を続ける。また、エネルギーとエントロフィは保存される。ケース (2) では、時間発展させると二つの渦はフィラメントを放出しながら並進し、微細な渦が残る。ケース (3) ではケース (2) と同じように最初にフィラメントを放出しながら並進するが、時間発展によって最後にフィラメントがなくなる（図4：(a, h) = (0.8, 0.4)）。

EMMで予測される大きな渦の運動や特異なふるまは全く現れなかった。

図5 Counter-rotating 渦の相互作用

4. 結論

楕円体モデルとCASL法の数値シミュレーションによって渦対の運動を調べた。ケース (1) におけるEMMの予測とCASL法の数値計算結果は一致した。しかし、他のケースの中で、EMMは渦が引き伸ばされて倒れる特異現象や大きな渦の運動を予測するか、数値計算では、渦はフィラメントを放出し、安定に運動する。Dritschel 純が提案するようなEMMの改良が必要とされるようである。
直線直角座標系における高分解能の大気局地数値モデルの開発

余偉明
東北大学大学院理学研究科地球物理学専攻

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、O（100 m）の水平方向分解能で大気地域数値モデルの実行が現実化することになると想定される。急峻な地形および複雑な物体をより適切に扱うため、適応的な数値手法を提案し、安定、効率的および正確な力学フレームの開発を行う。

2. 概要

我々は、次世代の大気局地数値モデルの開発を目的として行った。これは、地球の表面にある急峻な地形および複雑な物体を、より高い分解能で適切に扱うことが期待されるものである。本研究では、非定常三次元可圧縮ナヴィエ・ストークス方程式（Navier-Stokes equations）を計算するためのSIMPLER（Semi-Implicit Method for Pressure-Linked Equation Revised）アルゴリズムを用い、有限体積法（finite volume method）を用いた。我々は、慣例となっている地形追従規格化（terrain-following normalization）は行わずに、高さを垂直座標として用いるデカルト座標を採用した。地球の平均海面高度より上にある急峻な地形および全ての複雑な物体を扱うために、ブロックイング・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。ここで乱流計算のテストとして、都市建物群内の流れについて本モデルを実行した。シミュレーションの結果は、今後の大気局地数値モデル開発において本研究によって提案された数値手法の適用可能性を示すものであった。

3. 成果

図1は、仮想都市建物群の形状及び計算格子を示している。図2に建物群内の地表近くの水平速度が示されている。図には、物体の角周辺に数値不安定による流れの発生は見られず、シミュレーションの結果は予想したものであったので、この力学フレームに関して有望なものであると言える。今後の課題として実際の都市建物内の流れに関する計算を進める予定である。

図1 仮想都市建物群の形状及び計算格子
図2 建物群内の地表近くの水平速度分布
温度成層した液相反応乱流場での化学反応に及ぼす浮力効果

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2 国立環境研究所流域環境管理研究プロジェクト

1. 背景と目的
現実の乱流場に適用可能な数値計算手法としてLarge-Eddy Simulation (LES) が挙げられる。LESでは計算格子幅を大きさを設定することが可能である一方、計算格子より小さいスケール(Sub-grid Scale: SGS)に対しては適切なSGSモデルを与えなければならない。そのため、化学反応を伴う乱流場に対してLESを適用するためには、適切なSGS反応モデルを与える必要がある。近年、我々は温度成層が存在しない液相乱流場における反応モデルを開発し、その精度を実験結果と比較することにより確めた。しかし、開発された反応モデルが温度成層が存在する液相反応乱流場に対しても適用可能であるかどうかを、液相混合層と液相格子乱流場という代表的な二種類の乱流乱流場に対してLESを実行し、実験結果と比較検討することによって明らかにすることを目的とした。
そこで、本研究では、開発された反応モデルが温度成層が存在する液相反応乱流場においても適用可能であるかどうかを、液相混合層と液相格子乱流場という代表的な二種類の反応乱流場に対してLESを実行し、実験結果と比較検討することによって明らかにすることを目的とした。
一方、温度成層が存在する場合には浮力がSGSの流れ場に直接影響を及ぼす可能性が考えられる。本研究では、浮力がSGSの流れ場に及ぼす影響を考慮した場合と考慮しない場合の両方でLESを実行することにより、浮力がSGSに及ぼす影響を考慮する必要があるかどうかを明らかにした。

2. Large-Eddy Simulation
LESを実行するに当たっては、適切なSGSモデルが必要である。本研究ではSGSレイノルズ応力、SGS乱流物質フラックスおよびSGS乱流熱フラックスに対してDynamic SGSモデルを用いた。また、化学反応を計算するためには、適切なSGS反応モデルが必要である。これまでに我々は、SGSの濃度に影響を及ぼすことでより液相反応乱流場に対しても適用可能なSGS反応モデルを提案した。本研究でもこの反応モデルを用いた。

計算条件は実験と同じとした。混合層の場合には上層の流入速度を0.165 m/s、下層の流入速度を0.085 m/sに設定し、格子乱流場の場合にはテストセッション入口部に格子間隔(M=0.02 mの乱流格子を設置し、上下層の流入速度をともに0.18 m/sとした。温度成層が存在する場合には上下層の温度差を10Kとした。

3. 結果
図1(a)および図1(b)に液相混合層および液相格子乱流場における反応生成物濃度の鉛直方向分布をそれぞれ示す。なお、図1(a)はx=0.40 mの断面における結果、図1(b)はx/M=20の断面における結果である。図1(a)，(b)より、若干の差異は見られるものの、本LESの結果は実験結果とよく一致することがわかる。

4. 結言
本研究では、液相混合層と液相格子乱流場という代表的な二種類の乱流場に注目し、温度成層した液相反応乱流場に対してLESを実行し、その結果を実験結果と比較した。その結果、液相乱流場であっても浮力がSGSに及ぼす影響を考慮する必要がないこと、本LESは液相反応乱流場における乱流混合反応に及ぼす温度成層効果を正しく予測できることがわかった。
Figure 1  Vertical distributions of the mean concentration of chemical product (a) at $x=0.40$ m in the mixing layer and (b) at $x/M = 20$ in grid-generated turbulence. $y_m$ is defined by the vertical position with the mean concentration of species A equivalent to the cross-sectional mean concentration of species A.
衛星データと統合型数値モデルの融合による
釧路湿原の乾燥化現象の再現計算

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1. 背景と目的
近年、北海道の釧路川流域において、農地拡大のための平野部の森林伐採と河道直線化及び湿原の埋立に伴って、その下流流域にある釧路湿原では流れ込む土砂量の増加のために湿原の乾燥化・減少化が進行し従来の生態系が大きく変化するとともに、その解明及び対策が急がれている。著者らは釧路川流域流域(流域面積: 2204.7km²、釧路湿原を含む)を対象として気象ステーション(代表的な土地被覆である、湿原・牧草地・森林)3地点、地下水位30地点、河川水位13地点及びポーリング6地点から構成される現地観測ネットワークの確立を行った。これらの現地観測データ、様々な既存データ(数値標高データ、土地被覆、地質構造、粒度組成、地下水定水頭等)、及びMODIS衛星データを同化した、地上から地下までの水・熱移動の再現可能なモデル開発を行った。

2. 概要
grid型の川河流出モデル・衛星データを同化した陆面過程モデル・土壤水分方程式へのRichards方程式の導入による不飽和領域の鉛直方向へのメッシュの細分化(根による吸収などを考慮)及び3次元地下水モデルから構成される、地上から地下までの全ての領域での水・熱・物質移動の再現可能なプロセスベースの統合型数値モデルである、NICEモデル(NIES Integrated Catchment-based Eco-hydrology model)の開発及び改良を行った(図1)。それらのサブモデルは、土地被覆及び植生変化に大きく依存するためにこれらを対象と表現するモデルが必要であり、本モデルでは被覆面積指数(LAI)及び光合成有効放射率(FPAR)のMODIS衛星データから植生状態の時間的・空間的変化を考慮した。

3. 成果
図1 プロセスベースの統合型数値モデル（NICEモデル）の概念図

2で説明したNICEモデルを用いて、国立環境研究所のサーバーコンピュータ(NEC, SX-6)を用いた大規模モデルシミュレーションを行った。計算領域は釧路湿原及び釧路川流域(釧路市他4町1村、流域面積: 2204.7km²)を含む東西50km及び南北80kmの領域で、東西100×南北160×鉛直20のメッシュに分割した。また、河道は約320個の河道区分に分割した。計算ステップは、土壌水分量及び地下水位の推定の際には1時間、河川流量の推定の際には計算が安定になるように必要に応じて8〜30秒に断階変化させた。モデル計算結果はこれらの現地観測データと良好に一致し(地中温度・土壌水分量・地下水位・河川流量など)、NICEモデルは土地被覆・土壤構造、及び植生分布の時間的・空間的変化とそれに伴う
流域での水・熱収支の相互通用を考慮した非常に精度の高いものであることが明らかになった（Nakayama et al., 2003）。

以上をもとに、過去から現在までの土地利用データを利用してNICEモデルによるシミュレーションを行い、湿原の乾燥化が顕著な夏季における湿原域での再現計算を行った（図2）。3通りのシミュレーションには全て同一のForcingデータ、土壤データ、地質データを使用し、土地利用のみを変化させた。モデル計算に使用した土地利用データでは湿原内におけるヨシ・スゲ群落等の湿原固有種と近年侵入が著しいハンノキ群落との区別が明確でなく、ともに湿地という土地利用で分類されているため、湿原域での正確なシミュレーションは不可能であるが、流域の土地利用変化に伴う湿原への大まかな影響評価を行うことは可能であると考えられる。同図より、流域における森林伐採や農地化によって涵養量が減少し、その背後の湿原内部でも微小ではあるか地下水位が低下することになる（図2中の丸線で囲んだ領域）。

また、農地化された湿原の再湿地化や河道の再蛇行化によって地下水位は上昇し、水の滞留時間の増大による洪水遅延・下流への洪水被害の軽減、土砂捕捉による水質浄化が促進されると考えられている。今後、NICEモデルのような精緻な数値モデルと様々な政策シナリオを融合させることによって、短期的及び長期的時間スケールでの湿原生態系の回復手段の提示を行うことが非常に重要になってくると考えられる。
Program of
the 11th Supercomputer Workshop

October 21, 2003
National Institute for Environmental Studies
Tsukuba, Ibaraki, Japan
The 11th Supercomputer Workshop  
11:00～17:30, October 21, 2003  
National Institute for Environmental Studies, Tsukuba, Japan

11:00～11:05 Opening Address  
Shuzo Nishioka (Executive Director, Center for Global Environmental Research (CGER), National Institute for Environmental Studies (NIES))

11:05～11:15 Introduction of Research Programs  
Yasumi Fujinuma (CGER/NIES)

11:15～11:25 Simulation of Radiation Environment in Forest  
Tomomi Takeda (NIES), et al.

11:25～11:35 On the Stratospheric Chemical Transport Model in Meteorological Research Institute  
Kiyotaka Shibata (Meteorological Research Institute), et al.

11:35～11:45 The High-resolution Numerical Model of Heat Island Phenomena  
Yasunobu Ashie (Building Research Institute), et al.

11:45～11:55 Study on the Advancement of Troposphere Aerosol Process  
Masaru Chiba (Meteorological Research Institute), et al.

11:55～12:00 Lunch

13:00～13:15 Daily Runoff Simulation by an Integrated Catchment Model in the Middle and Lower Regions of the Changjiang Basin, China  
Seiji Hayashi (NIES), et al.

13:15～13:30 Ecosystem Modeling in East China Sea  
Masataka Watanabe (NIES), et al.

13:30～13:45 The Effects of Swell on the Mass Transfer Across the Air-sea Interface  
Satoru Komori (Kyoto University)

13:45～14:00 Design and Development of Atmospheric General Circulation Model for Terrestrial Planets: Basic Tests with a Spherical Shallow Water System  
Masatugu Odaka (Hokkaido University)

14:00～14:15 A Basic Experiment on Dependency of the Atmosphere Response to Various Distributions of Warm Tropical SST Anomaly in an Aqua-planet CGM  
Yoshi-Yuki Hayashi (Hokkaido University), et al.

14:15～14:30 Development of an Urban Meteorological Numerical Model in Cartesian Coordinate  
Sha Weiming (Tohoku University)

14:30～14:45 Generation of Cyclones by a Large-scale Mountain and Effects of Stratification and Rotation on It  
Yu Hozumi (Kyoto University), et al.

14:45～15:00 Scenario Dependence of Global Warming Projected by the MRI-CGCM2.3  
Akira Noda (Meteorological Research Institute), et al.

15:00～15:15 coffee break

15:15～15:30 Long-term Response of the Thermohaline Circulation to Climate Changes: a Theoretical Aspect  
Murakami Shigennori (Meteorological Research Institute), et al.

15:30～15:45 Analysis of Climate Change by the Aerosol Direct and Indirect Effects with a Global Three-dimensional Aerosol Transport-radiation Model  
Toshihiko Takemura (Kyushu University), et al.

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15:45〜16:00  Natural Variability of the Activity of Cumulus Convection Emerging in a Very Large Domain Two-dimensional Model
  Kensuke Nakajima (Kyushu University)

16:00〜16:15  Study of the Interannual Variability in Global Carbon Cycle with Parallel Atmospheric Transport Model
  Shamil Maksyutov (Frontier Research System for Global Change), et al.

16:15〜16:30  Radiative-convective Equilibrium with a Cloud Resolving Model
  Masaki Sato (Saitama Institute of Technology)

16:30〜16:45  Ozone Destruction Effects of Bromine Species
  Hideharu Akiyoshi (NIES), et al.

16:45〜17:05  Climate Change Simulations on the 20th Century with the CCSR/NIES/FRSGFC AGCM
  Toru Nozawa (NIES), et al.

17:05〜17:15  Supercomputer System of NIES
  Kuniyasu Hamada (HPC Center, HEC Group, NEC)

17:15〜17:30  Discussion

17:30  Closing Address
  Gen Inoue (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11回）プログラム

日時：平成15年10月21日（火）11:00～17:30

場所：独立行政法人 国立環境研究所 地球温暖化研究棟 交流会議室

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

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

11:15～11:25 森林の放射環境シミュレーション
○栄田 知己・小熊 宏之（国立環境研究所地球温暖化研究プロジェクト）

11:25～11:35 気象研究所化学輸送モデル（MJ99-CTM）を用いた中緯度における長期オゾン変動の解析と変動要因の解明に関する研究
○柴田 清孝・出川 眞・関山 剛（気象研究所環境・応用気象研究所）

11:35～11:45 ヒートアイランド数値モデルの高解像度化に関する研究
○足永 靖信1・一ノ瀬 俊明2・平野 佑二郎2・河野 孝昭1（1独建築研究所環境研究グループ、2国立環境研究所地球環境研究センター）

11:45～11:55 対流圈エアロゾル及びオゾン過程の高度化に関する研究
○尾葉 長・柴田 清孝（気象研究所環境・応用気象研究所）

11:55～13:00 昼休み

13:00～13:15 統合型流域モデルによる長江中下流域における日単位流出シミュレーション
○林 憲二・村上 正吾・中山 慎毅・徐 開鈦・渡辺 正孝（国立環境研究所流域環境管理研究プロジェクト）

13:15～13:30 東シナ海における水塊形成と長江河口域での塩分過遅上シミュレーション
○渡辺 正孝1・越川 海2・徐 開鈦2・村上 正吾2・石川 裕二2（1国立環境研究所水土壤環境研究領域、2国立環境研究所流域環境管理研究プロジェクト、3有水研開発コンサルタント）

13:30～13:45 大気・海洋間の物質輸送速度に及ぼすうねりの効果
○小森 悟（京都大学大学院工学研究科）

13:45～14:00 地球型惑星大気循環モデルの設計と開発：球面浅水方程式を用いた基礎的検討
○小高 正嗣1・石渡 正樹1・林 祥介1・竹広 真一1（1北海道大学大学院理学研究科、2北海道大学大学院地球環境科学研究所、3京都大学数理解析研究所）

14:00～14:15 水惑星GCMの熱帯域におかれたさまざまな暖水域分布に対する大気応答の依存性に関する基礎的実験
○林 祥介1・石渡 正樹1・小高 正嗣1・山田 由貴子1・中島 健介1（1北海道大学大学院理学研究科、2北海道大学大学院地球環境科学研究所、3九州大学理学研究所）

14:15～14:30 Development of an urban meteorological numerical model in Cartesian coordinate（直角直線座標系を用いた都市スケール気象数値モデルの開発）
○余 健昭（東北大学大学院理学研究科）

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14:30〜14:45 大規模山岳における低気圧性渦の剥離 ～カルマン渦の生成～
松尾 祐一・植田 洋匡・余 偉明（京都大学防災研究所、東北大学大学院理学研究科）

14:45〜15:00 新排出シナリオに基づく新しい気候変動シナリオの推計に関する研究
野田 彰・行本 誠史・内山 貴雄（気象研究所気候研究部）

15:00〜15:15 休憩

15:15〜15:30 気候変化に対する熱堆降の長期的応答 ～理論的側面～
村上 茂雄・鬼頭 昭雄・行本 誠史・野田 彰（気象研究所気候研究部）

15:30〜15:45 全球3次元自作輸送・放射モデルを用いた気候変動の解析
竹村 俊彦・野澤 徹・江守 正多・中島 麦至・日暮 明子（九州大学応用力学研究所、国立環境研究所大気圏環境研究領域、地球フロンティア研究システム、東京大学気候システム研究センター）

15:45〜16:00 大領域2次元積雪模型の自然変動
中島 健介（九州大学大学院）

16:00〜16:15 Study of the interannual variability in global carbon cycle with parallel atmospheric transport model（大気輸送モデルによる炭素循環の年々変動の研究）
Shamil Maksyutov1, P. K. Patra1, M. Ishizawa1, G. Inoue1, T. Nakazawa1,2,3,4,5
1Frontier Research System for Global Change, 2Center for Global Environmental Research, National Institute for Environmental Studies, 3Tohoku University

16:15〜16:30 霧解像モデルによる放射対流流柵
佐藤 重樹（埼玉工業大学）

16:30〜16:45 臭素系物質のオゾン破壊に及ぼす影響
秋吉 英治・今村 彰史・黒川 純一・渡川 雅之・菅野 誠治・中根 弘昭・高橋 正明（国立環境研究所地球環境研究プロジェクト、富士通FIP、地球フロンティア研究システム、国立環境研究所大気圏環境研究領域、東京大学気候システム研究センター）

16:45〜17:05 CCSR/NIES/FRSGC 大気オゾン模型を用いた20世紀の気候再現実験
野沢 徹・永島 達也・横畑 徳太・小倉 知夫・岡田 直知・竹村 俊彦・江守 正多・西村 照幸・木本 昌秀（国立環境研究所大気圏環境研究領域、九州大学応用力学研究所、地球フロンティア研究システム、東京大学気候システム研究センター）

17:05〜17:15 スーパーコンピュータのプログラムチューニング実績/事例
浜田 邦靖（NEC/HPCグループ/HPCセンター）

17:15〜17:30 総合討議

17:30 閉会挨拶
井上 元（国立環境研究所地球環境研究センター総括研究管理官）
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