

Development of a novel scheme for cloud and precipitation in a global aerosol-climate model with satellite observations

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<https://doi.org/10.15017/1866341>

出版情報 : 九州大学, 2017, 博士 (理学), 課程博士
バージョン :
権利関係 :



Development of a novel scheme for cloud and precipitation in a global aerosol-climate model with satellite observations

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September 2017

Development of a novel scheme for cloud and precipitation in a global aerosol-climate model with satellite observations

presented by

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A dissertation submitted to the
Department of Earth System Science and Technology
Kyushu University

for the degree of *Doctor of Sciences*

August 25, 2017

accepted on the recommendation of

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Abstract

Atmospheric aerosols have an important role in the climate system via their radiative scattering and absorption properties (aerosol-radiation interaction) as well as interactions with clouds (aerosol-cloud interactions). The aerosol-cloud interaction remains elusive and is one of the most uncertain processes in general circulation models (GCMs) via its nonlinear complexity. Almost all of the current GCMs represent the aerosol effects on cloud and precipitation microphysics based on empirical formulas and simplified assumptions, e.g., diagnostic treatment of rain (DIAG). In this respect, the parameterization of the aerosol-cloud-precipitation interaction must be elaborated at fundamental process-levels.

In this dissertation, firstly the aerosol effects on water cloud microphysics and macrophysics are investigated, using CloudSat and MODIS satellite observations over midlatitudes (from East Asia to the North Pacific). The systematic differences are found in the aerosol-induced modulation in drizzle intensity and in the vertical structure of the cloud-to-rain conversion process, among regional and seasonal scales. The cloud and precipitation microphysics are also susceptible to the atmospheric stability, as well as the aerosol loading. These results suggest that the aerosol-cloud interactions in the real atmosphere depend not only on the cloud microphysics but also the cloud types, thermodynamics, and surrounding meteorological field (microphysics-dynamics interactions), which are not considered in practically all of the current GCMs.

Next, the performance of autoconversion schemes (mass transfer from cloud water to rainwater by the coalescence of cloud droplets) parameterized in GCMs is evaluated to understand the potential uncertainty in the aerosol-cloud interactions. Compared to satellite observations, a poor representation of cloud macrophysical structure and optically thicker low cloud are found in the simulations with any autoconversion schemes. As a result of the cloud-radiation interaction, the difference in the global mean net cloud radiative forcing (NetCRF) among the schemes reaches 10 W m^{-2} . The discrepancy between the observed and simulated NetCRF is especially large in a high liquid water path (LWP) regime. The potential uncertainty in the parameterization of the autoconversion process is non-negligible, but none of the tested schemes significantly improve the bias in the cloud radiative effect yet. This means that more fundamental errors are still left in the other processes of the model.

To understand the fundamental model uncertainties suggested above, the response of LWP to aerosol perturbations is investigated from the perspective of GCM and the A-Train remote sensing. The model exhibits an increase in LWP with increasing aerosol loading almost over the

globe, while the sign of the response of the LWP from the A-Train varies from region to region. The satellite-observed response of the LWP is closely related to meteorological and/or macrophysical factors, in addition to the microphysics. The model does not reproduce this variability of cloud susceptibility because the parameterization of the autoconversion process assumes only suppression of rain formation in response to increased cloud droplet number concentration N_c , and does not consider macrophysical aspects that serve as a mechanism for the negative responses of the LWP via enhancements of evaporation and precipitation. Model biases are also found in the precipitation microphysics, which suggests that the model generates rainwater readily even when little cloud water is present. This essentially causes projections of unrealistically frequent and light rain, with high cloud susceptibilities to aerosol perturbations.

In order to overcome the above deficiencies, an advanced cloud-precipitation scheme is developed, and implemented into an aerosol-climate model called MIROC-SPRINTARS. The new scheme features prognostic treatment of rain (PROG) based on a two-moment bulk representation. Sensitivity experiments are performed using a single-column model for a specific warm rain case, which suggest the best combination of the autoconversion scheme, raindrop size distribution, and sub-time step for microphysics. Both rain mass and number mixing ratios are now stored across multiple time steps, and controlled strictly by the sub-time stepping iteration (~ 60 sec or less) while the standard model time step is 10–20 min approximately.

The performance of the new scheme is evaluated by global simulations using satellite datasets (e.g., CloudSat, MODIS, CERES, SSMI, GPCP, and ISCCP). In many respects, the PROG scheme improves model biases in the standard DIAG scheme. For example, the relative contribution of accretion is significantly enhanced against that of autoconversion, whose balance is consistent with observations. The horizontal and vertical distributions of simulated rainwater are also consistent with the CloudSat observations and a cloud resolving model simulation. It should be emphasized that shallow drizzle, which is frequently observed in coastal stratocumulus deck, is successfully simulated in the new scheme. Consequently, the bias in shortwave cloud radiative forcing, often called “too bright low-cloud problem”, is significantly reduced, in good agreement with the CERES observation.

Furthermore, both increasing and decreasing responses of LWP to aerosol perturbations are found in the new PROG scheme as observed in the A-Train satellites and large-eddy simulations in previous literatures, while the traditional DIAG approach shows only increasing LWP response of the so-called cloud lifetime effect. This “bidirectional” LWP response appeared in the new

scheme is robustly related to the cloud life stage. These improvements support that the prognostic precipitation framework developed in this thesis has great advantages in process representations of not only cloud-precipitation microphysics and cloud radiative forcing, but also cloud system responses including the “buffered system” behaviors.

The findings and progress that have been presented in this thesis will contribute significantly to more accurate climate predictions because the main source of uncertainties in climate modeling stems from the representation of cloud and precipitation.

This dissertation partly included the published articles by the author:

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2. © 2015 T. Michibata and T. Takemura: American Geophysical Union (Chapter 3)
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1 General Introduction

Earth’s climate system is composed of five major components or subsystems: atmosphere, hydrosphere, biosphere, land surface, and cryosphere, which mutually interact with each other. The recent development of a human activity significantly alters the environment and climate, which will be one of the traumatic experience in the long history of the Earth, while the estimates of anthropogenic forcing include significant uncertainty. In this thesis, we turn our attention to the atmosphere and its complex interaction in the climate systems (atmosphere–anthroposphere–hydrosphere interactions).

The feedback process among aerosols, water vapor, cloud, and precipitation remains elusive, and is one of the largest uncertainties in climate modeling (e.g., Lohmann and Feichter, 2005; Stevens and Feingold, 2009; Seinfeld et al., 2016). Atmospheric aerosols play significant roles on climate directly and indirectly, so-called “aerosol-radiation interactions (ARI)” (e.g., Coakley et al., 1987; Charlson et al., 1992) and “aerosol-cloud interactions (ACI)” (e.g., Twomey, 1977; Albrecht, 1989), respectively. The former process (ARI) depends strongly on the scattering and absorption properties of aerosols themselves; dark aerosols (e.g., black carbon) contribute to heating atmosphere, whereas bright ones (e.g., sulfate) operate to cooling. This includes adjustment by absorbing aerosols, such the cloud burning (Ackerman et al., 2000; Hansen et al., 1997), which causes local heating and stability modification semidirectly (Jacobson, 2001). The latter process (ACI) is related to the aerosol number concentration N_a . An increase in aerosols supplies more cloud condensation nuclei (CCN), which leads to more numerous smaller cloud droplets and thus brighter cloud. This is referred to as the “cloud albedo effect” (Twomey, 1977). Furthermore, rain formation is susceptible to the aerosol (and/or CCN) number because the onset of precipitation is suppressed due to inefficient collision-coalescence by smaller cloud droplets. This aerosol induced modulation is referred to as the “cloud lifetime effect” (Albrecht, 1989).

The fifth assessment report (AR5) of the Intergovernmental Panel on Climate Change (IPCC) described that the total of the effective radiative forcing by the ARI (ERF_{ari}) and ACI (ERF_{aci}) is approximately -0.90 W m^{-2} , which is comparable to the ERF by anthropogenic carbon dioxide (CO_2) of $+1.82 \text{ W m}^{-2}$. Although the ERF_{ari} and ERF_{aci} are almost the same value (-0.45 W m^{-2}), the uncertainty in the ERF_{aci} is larger than that in the ERF_{ari}. To estimate the magnitude of the ACI is difficult both from observations and modeling, since the spatial/temporal

variability of aerosol and cloud is quite large. The meteorological factors also influence aerosol and cloud fields simultaneously, which would incur the spurious correlation of the ACI due to their covariance (e.g., Grandey and Stier, 2010; Sorooshian et al., 2013; Gryspeerdt et al., 2014).

One of the key indices to quantify the magnitude of the ACI is the response of cloud liquid water path (LWP) to aerosol perturbations (Wang et al., 2012; Chen et al., 2014). It is noteworthy that there are two competing pathways of LWP in literature, that reported both increases (traditional “cloud lifetime” hypothesis) and decreases (recent “buffered system” hypothesis) in LWP to increased aerosols. Most of general circulation models (GCMs) show positive relationship between LWP and N_a (or CCN) since the models reduce the autoconversion rate (mass transfer from cloud water to rainwater) for incremental changes in the cloud droplet number concentration N_c assuming lifetime effect (Quaas et al., 2009; Wang et al., 2012). On the other hand, some observational studies (e.g., Matsui et al., 2006; Lebsock et al., 2008; Chen et al., 2014) and high-resolution process modeling (e.g., Ackerman et al., 2004; Small et al., 2009; Seifert et al., 2015) have demonstrated the “anti”-lifetime response, namely, negative relationship between LWP and aerosol proxies. This competing negative response is now considered as the “buffered system” (Stevens and Feingold, 2009), which mitigates the aerosol perturbations and drives negative feedback on the system scale. Despite the importance in an accurate climate simulation, the exact mechanism is still unclear why two competing responses appear and how they depend on meteorological factors.

GCMs also have two major issues on the cloud-precipitation system. The one is “too few too bright low-cloud problem” (Nam et al., 2012), and the other is “too fast too frequent rain formation” (Suzuki et al., 2015). Nam et al. (2012) found that the models overestimated shortwave reflectance even though the cloud cover was consistent with observations. This is reduced by error compensations that underestimate low-cloud cover and/or overestimate the occurrence of high-clouds above the low-clouds. These tendencies of GCMs were also found by Wang and Su (2013) and Calisto et al. (2014). Suzuki et al. (2015) evaluated the cloud-to-rain water conversion process through multi-model analysis, and reported that GCMs tended to generate rainwater too fast and too frequently compared to satellite observations. The both issues in GCMs described above suggest that the fundamental flaw exists in their process representation. More physically based treatment of microphysics is required to reproduce the nonlinear system in aerosol-cloud-precipitation interactions (Seinfeld et al., 2016).

A possible solution to overcome the above issues is to elaborate the treatment of rain in the

model. Almost GCMs treat precipitating hydrometeors (i.e., rainwater and snowflake) diagnostically, which assumes all of diagnosed rainwater precipitate to the surface within a single model time step (e.g., Ghan and Easter, 1992). This is a useful approach especially for large hydrometeors, whose horizontal advection and vertical sedimentation are not important in GCMs with coarse spatial/temporal resolutions. For example, assuming a raindrop with 5 m s^{-1} fall speed, it falls through 6 km during 20 min which is a typical time step in GCMs. Although the diagnostic treatment of rain contributes to reduce a computational cost, small raindrops (drizzle) which do not reach the surface within a single time step are ignored. Stratocumulus system frequently forms drizzle with occurrences of 20–50 % (e.g. Leon et al., 2008; Wu et al., 2015), which is quite important for shortwave reflectance (Muhlbauer et al., 2014) and the onset of precipitation (Wood, 2005b; vanZanten et al., 2005), thus hydrological lifecycle (Wood, 2012).

Recently, advanced cloud-precipitation framework was developed by adding prognostic equations for precipitation (Gettelman et al., 2015; Sant et al., 2015). The prognostic treatment of rain can keep the memory of rainwater in the atmosphere across multiple time steps, which enhances the relative contribution of accretion over autoconversion (Posselt and Lohmann, 2008; Gettelman et al., 2015) in good agreement with in-situ measurements (Wood et al., 2011; Wood, 2005a; Gettelman et al., 2013). Sant et al. (2015) demonstrated the improved representation of vertical structures of cloud and precipitation hydrometeors applying prognostic treatment of rain, which is also one of the critical issues in GCMs (e.g. Jiang et al., 2012; Komurcu et al., 2014; Li et al., 2012; Cesana et al., 2015). Since the liquid/ice phase partition and their vertical structures directly influence on the global radiation budget (Matus and L’Ecuyer, 2017; Cesana and Storelvmo, 2017), the introduction of prognostic rain scheme would be a fruitful concept on fundamental model development.

This thesis is organized as follows: Chapter 2 discusses the differences in the aerosol-cloud interactions between land and ocean, as well as those between JJA and DJF, over the East Asia and North Pacific based on satellite observations. In Chapter 3, sensitivity experiments are carried out in a global aerosol-climate model with commonly used five autoconversion schemes to examine the potential biases. Chapter 4 unveils the source of discrepancies in aerosol-cloud-precipitation interactions between GCMs and satellite retrievals through process-oriented model evaluations. In order to overcome the biases between models and observations, an advanced two-moment bulk microphysics scheme including prognostic precipitation is developed in Chapter 5. The magnitude of aerosol-cloud interactions and LWP susceptibility are evaluated by employing both

the new prognostic and traditional diagnostic rain scheme in Chapter 6. Finally, the summary of contributions of this thesis and future work are described in Chapter 7.

2 Constraining Aerosol-Cloud Interactions using Satellite Data

Abstract

This chapter examines the characteristics of the microphysics and macrophysics of water clouds from East Asia to the North Pacific, using data from active CloudSat radar measurements and passive MODerate-resolution Imaging Spectroradiometer (MODIS) retrievals. The goals are to clarify differences in microphysics and macrophysics between land and oceanic clouds, seasonal differences unique to the midlatitudes, characteristics of the drizzling process, and cloud vertical structure. In pristine oceanic areas, fractional occurrences of cloud optical thickness (COT) and cloud droplet effective radius (CDR) increase systematically with an increase in drizzle intensity, but these characteristics of the COT and CDR transition are less evident in polluted land areas. In addition, regional and seasonal differences are identified in terms of drizzle intensity as a function of the liquid water path (LWP) and cloud droplet number concentration (N_c). The correlations between drizzle intensity and LWP, and between drizzle intensity and N_c , are both more robust over oceanic areas than over land areas. This chapter also demonstrates regional and seasonal characteristics of the cloud vertical structure. The results suggest that aerosol–cloud interaction mainly occurs around the cloud base in polluted land areas during the winter season. In addition, a difference between polluted and pristine areas in the efficiency of cloud droplet growth is confirmed. These results suggest that water clouds over the midlatitudes exhibit a different drizzle system to those over the tropics.

Keywords:

water cloud physics, drizzle, aerosol–cloud interaction, CloudSat

Key Points:

- The characteristics of drizzling process in the mid-latitudes were clarified.
 - Seasonal/regional differences in the cloud physics were examined.
 - Cloud microphysical vertical structure was sensitive to aerosols concentration.
-

2.1 Introduction

The Earth’s radiation budget is affected to some extent by the scattering and absorption properties of aerosol, which are referred to as aerosol–radiation interactions. In addition, aerosol particles play an important role in the climate system by serving as cloud condensation nuclei (aerosol–cloud interaction). This affects cloud optical thickness (COT) and cloud particle size (e.g., Twomey, 1977) as well as cloud lifetime (e.g., Albrecht, 1989). However, accurate and quantitative evaluation of these indirect aerosol effects is required to address the considerable uncertainty related to the heterogeneous nature of the spatial and temporal distributions of aerosols. With respect to numerical models, many climate models have been developed and improved for an accurate estimation of the global radiation balance. Practically all of the climate models, however, have uncertainty in their cloud precipitation parameterization schemes (e.g., Suzuki et al., 2013a) due to the difficulty of representing the complex aerosol–cloud interactions.

The cloud profiling radar (CPR) of CloudSat, whose mission began in 2006, may help clarify the details of cloud physical properties (Stephens et al., 2002), including vertical information that cannot be obtained from conventional satellite passive sensors, and is important in clarifying indirect aerosol effects. Research on the physical properties of water clouds has advanced significantly in the last few years. Haynes and Stephens (2007) studied the relationships between cloud thickness and precipitation in the marine tropics and found regional differences in the cloud vertical structure (shallow, middle, and deep modes) of precipitating clouds. Lebsock et al. (2008) investigated mainly aerosol–cloud interactions based on multi-sensor satellite observations and found a relationship between variations in the cloud liquid water path (LWP) and the thermodynamic conditions. Kubar et al. (2009) compared the physical properties of water clouds in regions over tropical and subtropical oceans and stressed the importance of cloud macrophysics and microphysics for drizzle frequency and intensity. They also investigated which parameters were important for drizzle processes, focusing on macrophysics (cloud thickness and LWP) and microphysics (cloud droplet effective radius (CDR) and cloud droplet number concentration (N_c)).

Sorooshian et al. (2009) performed a binning study of LWP to clarify the effects of aerosol perturbation (e.g., precipitation susceptibility, aerosol cloud interactions) and suggested that intermediate LWP ($\sim 500\text{--}1000\text{ g m}^{-2}$) cloud tends to be more susceptible to aerosol than shallow cloud with low LWP. Furthermore, they expanded the study of Stephens and Haynes (2007), who introduced a method of estimating conversion (from cloud water to rainwater) rates from CloudSat-CPR and MODerate-resolution Imaging Spectroradiometer (MODIS)-retrieved data

and discussed the relationships between conversion rate and aerosol types, associated with the category of lower-tropospheric static stability (LTSS) and LWP (Sorooshian et al., 2013).

Nakajima et al. (2010) and Suzuki et al. (2010) attempted to visualize the vertical structure of cloud on a global scale using a method that they termed “contoured frequency by optical-depth diagram” (CFODD). Kawamoto and Suzuki (2012) applied CFODD to investigate precipitation processes and demonstrated that precipitation over the Amazon occurs in optically thicker locations than is the case over China.

Many researchers have investigated the physical structures and precipitation characteristics of low-level water clouds based on satellite data, as mentioned above. However, most of these studies were limited to the tropics/subtropics or areas over oceans; only a few have compared clouds over land and ocean. Very few have focused on East Asia, where some areas have significant levels of air pollution (e.g., Kawamoto and Suzuki, 2013). Therefore, clouds in these regions may exhibit drizzle characteristics that differ from those of clouds over tropical oceanic areas.

This study focuses on seasonal differences in water clouds that are characteristic of the midlatitudes and compares the characteristics of clouds over China (a region with considerable anthropogenic aerosols) with those over the North Pacific (a clean/pristine environment). This chapter also analyze the transition processes of drizzle over both land and ocean (e.g., Nakajima et al., 2010) in the midlatitudes, which have been evaluated in only a few other studies.

2.2 Data and Methodology

2.2.1 CloudSat and MODIS

CloudSat, launched by the National Aeronautics and Space Administration (NASA) in 2006, was the first project to include a spaceborne millimeter-wavelength (3 mm, frequency = 95 GHz) radar (Stephens et al., 2008) to help resolve the vertical structure of cloud droplets. The vertical and spatial resolutions of the CloudSat data products are approximately 480 m and 1.4×1.8 km (across and along tracks), respectively. However, the data are vertically $2\times$ oversampled, and therefore ~ 240 m sampled data are available (Stephens et al., 2008). The visible COT and CDR near the cloud top were obtained from the 2B-TAU product (Polonsky, 2008), and also radar reflectivity and the cloud mask from the 2B-GEOPROF product (e.g., Mace et al., 2007; Marchand et al., 2008). Temperature and pressure data for each altitude were obtained from the European Center for Medium-Range Weather Forecasts Auxiliary (ECMWF-AUX) objective analysis (Partain, 2007). The analysis periods were June, July, and August (JJA) from 2007

to 2009, and December, January, and February (DJF) from 2006 to 2009 (i.e., December 2006–February 2009).

The passive sensor MODIS traverses aerosol–cloud properties at high frequency and resolution, using 36-channel spectral bands (Platnick et al., 2003; Remer et al., 2005). The level 3 (collection 5.1) $1^\circ \times 1^\circ$ gridded aerosol optical depth (AOD) at $0.55 \mu\text{m}$ from Aqua/MODIS (Parkinson, 2003), which is a part of the A-Train constellation (Stephens et al., 2002), is used in our study.

The following Eq. (2.1) was used to estimate N_c (e.g., Brenguier et al., 2000; Wood, 2006; Kubar et al., 2009),

$$N_c = \sqrt{2} B^3 \Gamma_{\text{eff}}^{1/2} \frac{\text{LWP}^{1/2}}{r_e^3} \quad (2.1)$$

where $B = (3\pi\rho_w/4)^{1/3} = 0.0620$, ρ_w is the density of liquid water, and Γ_{eff} is the adiabatic rate of increase in the liquid water content with height, which is a function of two variables, profile of temperature and pressure, as shown in Fig. 1 of Wood (2006). The difference in CDR retrieval error between land and ocean, e.g., due to the differences in cloud type (e.g., Zhang et al., 2012b), may also cause uncertainty in estimation of N_c . However, the present study apply a CDR uncertainty threshold of $< 1 \mu\text{m}$, as mentioned above, which reduces N_c uncertainty as much as possible. Other possible errors due to the assumption of deriving N_c (e.g., adiabaticity, vertical homogeneity) are documented elsewhere (e.g., Grandey and Stier, 2010; Kubar et al., 2009). In addition, LWP was calculated by the following Eq. (2.2) (Brenguier et al., 2000),

$$\text{LWP} = 5 \tau_c r_e / 9 \quad (2.2)$$

where τ_c and r_e were obtained from MODIS retrieval, matched along the CloudSat footprint (i.e., CloudSat 2B-TAU product, mentioned earlier).

2.2.2 Regions and Methods

Figure 2.1 shows maps of the regions investigated in this study. Inland includes the Gobi Desert. An area of northeastern China (NE China) was selected to study the effects of soil dust aerosols transported from the Gobi and Taklamakan Deserts. Human activity generates many anthropogenic aerosols in the Industrial area, and this region is one of the most air-polluted areas in the world (upper panel of Fig. 2.1). Some areas of Japan region also discharge anthropogenic aerosols, but the main reason for selecting this region is to compare it with the Industrial area. The outflow regions of anthropogenic aerosols are referred to as North Pacific 1, 2, and 3 in order

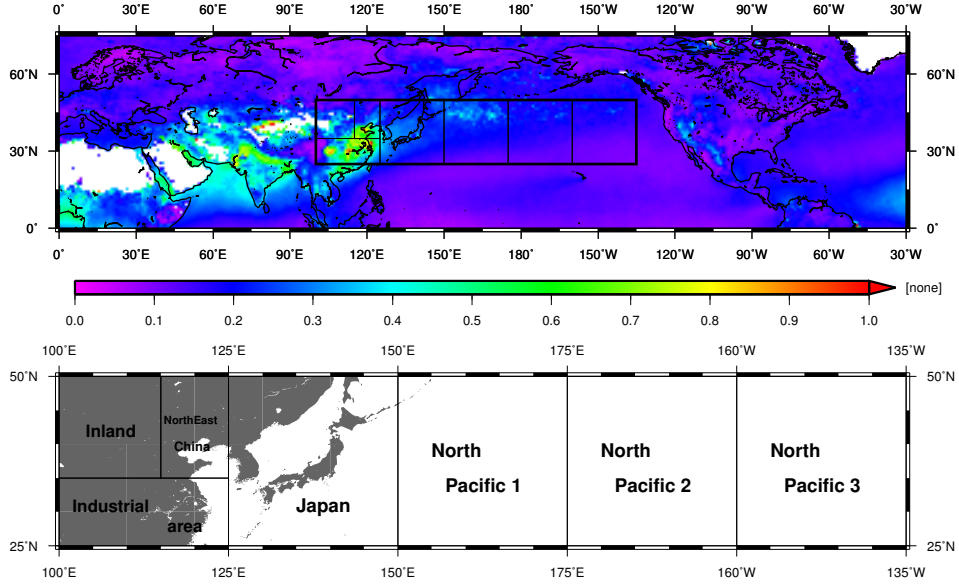


Fig. 2.1: Whole (top) and individual (bottom) regions in this study. Spatial distribution of aerosol optical thickness τ_a (550 nm) for the 3 year mean derived from monthly Aqua/MODIS level 3 products are illustrated in the top panel. Missing values are shown in white.

of their distance from East Asia. The main focus is to investigate how large amounts of aerosols transported from East Asia affect cloud properties in these areas.

This study focuses only on low-level water clouds, because most aerosols remain in the lower troposphere. Water clouds are defined as those with a cloud mask value greater than 30 (good/strong echo), which means high-confidence detection (estimated false detection $< 4.3\%$; see Marchand et al. (2008), Table 1), and a temperature above 273 K for the entire cloud layer. Furthermore, this study uses only the data with uncertainty values of less than 3 and $1 \mu\text{m}$ for COT and CDR, respectively. Multilayered clouds are excluded from the analyses to avoid ambiguous statistics.

LTSS is defined as the difference in potential temperatures between 700 hPa and the surface (Klein and Hartmann, 1993). This index was calculated from the ECMWF-AUX product (vertical temperature and pressure profiles).

2.3 Results

2.3.1 Cloud Physical Properties for Each Area

Table 2.1 lists the physical properties of clouds over each of the seven areas. DJF values are given in parentheses. The land–sea mask is not applied in our analysis, and therefore the data for

Table 2.1: Cloud physical parameters in each area. JJA and DJF values are 3 year means. DJF values are shown in parentheses. Maximum values are shown in bold and minimum values are underlined. Maximum radar reflectivity in the cloud layer ($Z_{\max}$) is used for precipitation categories (no precipitation; $Z_{\max} < -15$, drizzle; $-15 \leq Z_{\max} < 0$, rain; $0 \leq Z_{\max}$). The Inland and NE China regions in DJF, where none or few samples met the criteria, are shown by “not available (N/A)”.

	Land				Ocean		
	Inland	NE China	Industrial area	Japan	North Pacific 1	North Pacific 2	North Pacific 3
The number of samples	<u>693</u> (0)	1315 (1)	3927 (4540)	11914 (10118)	20674 (15920)	25029 (17455)	44064 (31949)
τ_a	0.29 (0.18)	0.40 (0.30)	0.49 (0.44)	0.23 (0.21)	0.24 (0.16)	0.17 (<u>0.14</u>)	<u>0.14</u> (<u>0.14</u>)
τ_c	22.2 (N/A)	24.5 (<u>0.73</u>)	19.5 (35.9)	22.0 (26.3)	19.9 (21.5)	17.9 (19.4)	<u>16.4</u> (18.7)
r_e [μm]	<u>11.9</u> (N/A)	<u>11.9</u> (<u>10.0</u>)	12.3 (10.5)	15.8 (14.5)	18.1 (17.8)	18.5 (18.0)	18.0 (17.3)
LWP [g m^{-2}]	148 (N/A)	161 (<u>4</u>)	<u>129</u> (205)	189 (207)	197 (215)	185 (197)	167 (180)
N_c [cm^{-3}]	154 (N/A)	139 (28)	125 (257)	77 (113)	51 (55)	42 (48)	<u>41</u> (50)
Maximum Z_e [dBZ_e]	-5.8 (N/A)	<u>-8.1</u> (<u>-27.3</u>)	0.8 (-1.1)	0.5 (0.9)	0.1 (2.0)	-0.3 (0.7)	-1.5 (-1.2)
[%] with no precipitation	67.1 (N/A)	70.5 (N/A)	49.6 (61.5)	46.2 (43.5)	<u>42.3</u> (<u>35.3</u>)	43.3 (40.1)	46.4 (45.5)
[%] with drizzle	28.1 (N/A)	<u>26.5</u> (N/A)	33.5 (<u>29.1</u>)	34.0 (34.7)	39.2 (38.5)	40.6 (38.5)	40.5 (40.9)
[%] with rain	4.8 (N/A)	<u>3.0</u> (N/A)	16.9 (<u>9.4</u>)	19.8 (21.8)	18.5 (26.2)	16.1 (21.4)	13.1 (13.6)
Cloud-top height [km]	3.7 (N/A)	2.7 (<u>1.4</u>)	3.4 (2.3)	2.4 (2.4)	2.1 (2.2)	1.9 (2.0)	<u>1.8</u> (1.7)
Cloud-base height [km]	2.7 (N/A)	1.6 (1.2)	2.3 (1.4)	1.2 (1.1)	<u>1.0</u> (<u>1.0</u>)	<u>1.0</u> (<u>1.0</u>)	<u>1.0</u> (<u>1.0</u>)
Geometrical thickness [km]	1.0 (N/A)	1.1 (<u>0.2</u>)	1.2 (0.9)	1.2 (1.3)	1.1 (1.3)	1.0 (1.0)	<u>0.8</u> (0.8)
LTSS [K]	13.8 (N/A)	15.2 (16.8)	<u>12.2</u> (19.6)	16.5 (15.9)	19.4 (<u>15.8</u>)	18.3 (16.8)	18.2 (17.5)

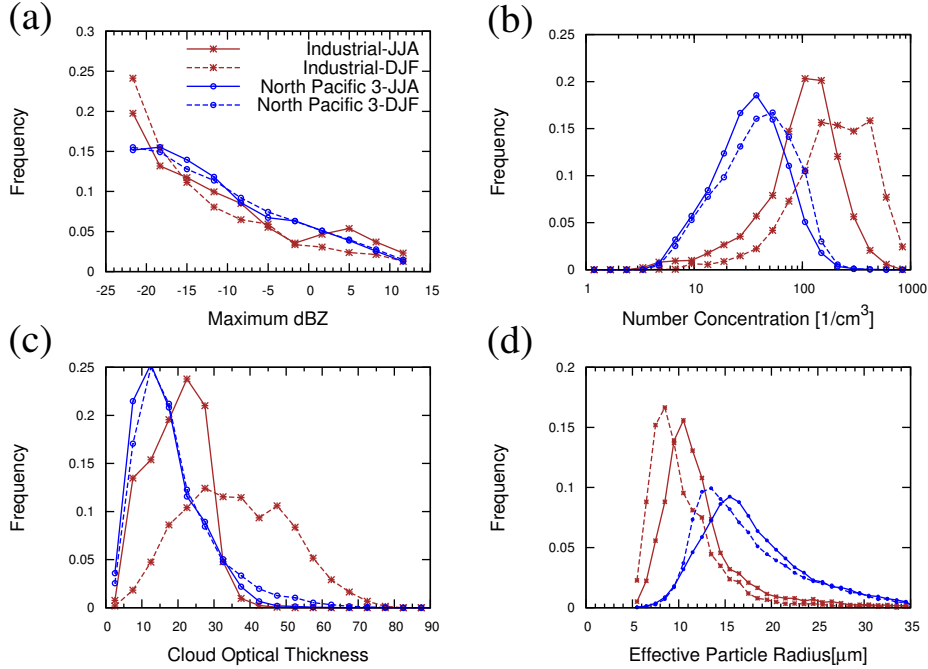


Fig. 2.2: Probability distribution functions (PDFs) of each cloud physical variable, (a) maximum radar reflectivity $Z_{\max}$ [dBZ_e], (b) cloud droplet number concentration N_c [cm^{-3}], (c) cloud optical thickness τ_c , and (d) cloud effective particle radius r_e [μm] for Industrial area and North Pacific 3 in JJA (solid line) and DJF (dotted line).

the Japan, NE China, and Industrial area, including the ocean part, do not necessarily represent data only over the continent. The results suggest that the precipitation occurrence is related to LWP (e.g., North Pacific 1 where higher LWP region is accompanied with high ‘[%] with rain’; Table 1), except for the Industrial area (i.e., high LWP but lower ‘[%] with rain’, and vice versa). It is noteworthy that there are large seasonal differences of more than 7 K in LTSS in the Industrial area. Therefore, there is a possibility of different cloud types over the Industrial area; i.e., cumulus cloud in JJA (unstable lower LTSS environment) than over the oceanic area. The passive MODIS sensor tends to retrieve errors on cumulus inhomogeneous cloud (e.g., Zhang et al., 2012b; Zhang and Platnick, 2011; Zinner et al., 2010) because of its simplifying assumptions; i.e., clouds are plane-parallel and homogeneous, any effects of drizzle/rain drops are ignored (Zinner et al., 2010), etc. These assumptions may lead to retrieval bias of CDR; e.g., shadowing effects can lead to underestimation of COT and overestimation of CDR (Marshak et al., 2006). The smaller COT and larger CDR are estimated with increasing cloud inhomogeneity, which results in underestimation of LWP for cloudy scenes (Painemal et al., 2013). Therefore, care should be taken with regard to this background of CDR retrieval error and underestimation of LWP, especially over the Industrial area in JJA.

Figure 2.2 shows the probability distribution function (PDF) of each cloud physical variable. The distribution of maximum radar reflectivity in the cloud layer ($Z_{\max}$) (Fig. 2.2a) is similar for both the Industrial area and North Pacific 3, although a slight shift to weaker $Z_{\max}$ was observed for the Industrial area. The tendencies of smaller CDR values, larger N_c values, and optically thicker clouds were observed over land areas than over the oceanic regions in Fig. 2.2 and Table 2.1, supporting the findings of previous studies (e.g., Kawamoto et al., 2001). However, these results are not as obvious in the region over Japan as in other land areas, Inland, NE China, or the Industrial area. It is possible that the properties of clouds over NE China are affected in a complex manner by dust aerosols from the adjacent western deserts and emissions of anthropogenic aerosols from highly populated areas, such as Beijing. The North Pacific 1 area has slightly larger values for COT, LWP, and N_c compared with the other oceanic areas, and the values of CDR are almost the same for all oceanic areas. Small seasonal differences are observed during JJA and DJF over the three oceanic areas; these differences are more obvious over the four land areas, which may be due to the high levels of aerosols in DJF, when atmospheric conditions are most stable.

The mode radii are approximately $15\ \mu\text{m}$ over the three oceanic areas, whereas they are ap-

proximately $9\text{ }\mu\text{m}$ over the Industrial area in DJF, which may result in less efficient precipitation. The following subsections discuss how differences in the physical properties of clouds over land and ocean regions affect the rainfall characteristics.

2.3.2 COT–CDR Diagram

COT and CDR are commonly considered cloud physical variables. The fact that the correlation between these parameters reflects cloud growth (only liquid phase warm cloud) and precipitation processes has been well documented in previous studies based on satellite observations (e.g., Nakajima et al., 1991; Nakajima and Nakajima, 1995). That is, both COT and CDR increase early in the growth process of cloud droplets, resulting in a positive correlation between them. The cloud particles grow to almost $15\text{ }\mu\text{m}$, and precipitation begins. With precipitation, COT decreases and CDR increases due to coalescence. This precipitation process leads to a negative correlation pattern. Suzuki et al. (2006) extended these analyses, and successfully simulated the pattern using a spectral-bin microphysics model. Suzuki et al. (2011) documented fractional occurrences as a function of COT and CDR for each rain category (no precipitation, drizzle, and rain), and compared A-Train observations with model simulations.

Figure 2.3 shows fractional occurrences on COT–CDR diagrams for each rain category ([A] no precipitation; $Z_{\text{max}} < -15$, [B] drizzle; $-15 \leq Z_{\text{max}} < 0$ and [C] rain; $0 \leq Z_{\text{max}}$) (Comstock et al., 2004; Stephens and Haynes, 2007). The diagrams in the pristine remote ocean (North Pacific 3, Fig. 2.3g–l) reveal that the main group systematically shifts from the lower COT–CDR region to the higher COT–CDR region with an increase in the rain category (i.e., from no precipitation to rain, with a monotonous increase in LWP and a slight decrease in LTSS) during both seasons. This tendency was also reported by Suzuki et al. (2011) and Kawamoto and Suzuki (2013). The fact that JJA (Fig. 2.3g–i) and DJF (Fig. 2.3j–l) have similar distributions suggests that the relation between COT and CDR has considerable universality with the rain categories over oceanic areas. However, in the Industrial area where air pollution by anthropogenic aerosols is severe, the transition pattern is not as clear as over the ocean, and the variations of LWP are relatively small. The category Rain in JJA (Fig. 2.3c) has relatively high values of fractional occurrence (approximately 0.2–0.5) in the small COT–CDR region ($\text{COT} < 15$, $\text{CDR} < 15\text{ }\mu\text{m}$), while most values in this region (see Fig. 2.3i, l) are less than 0.2. Furthermore, large numbers of samples were concentrated in this region and that the cloud-top height in the Industrial area is much higher (3.3 km) than that in the North Pacific 3 area (2.4 km). These findings suggest the

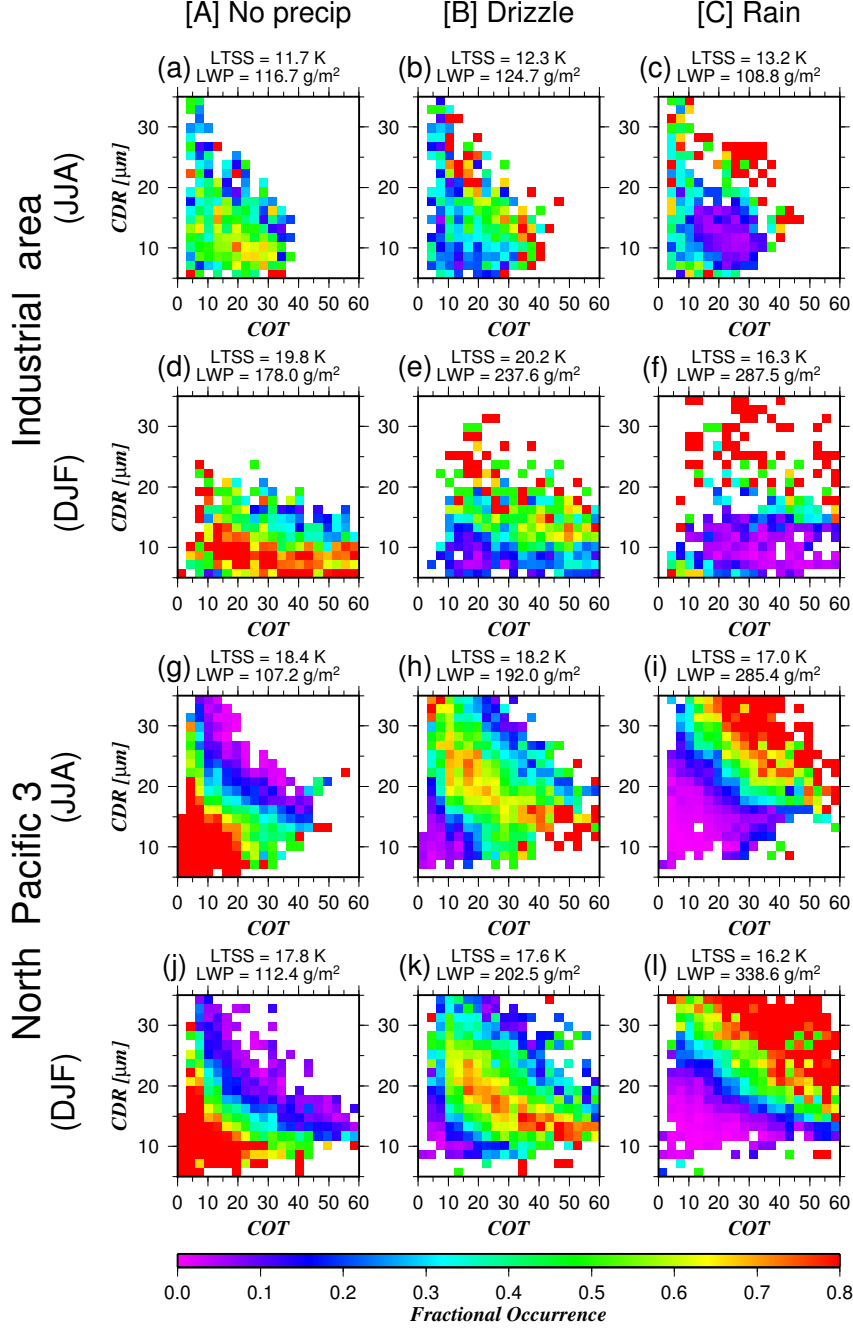


Fig. 2.3: Fractional occurrences of cloud optical thickness (COT) and cloud droplet effective radius (CDR) for each rain category: [A] no precipitation ($Z_{\max} < -15$), [B] drizzle ($-15 \leq Z_{\max} < 0$), and [C] rain ($0 \leq Z_{\max}$). (a–c) are for the Industrial area in JJA, (d–f) for the Industrial area in DJF, (g–i) for the North Pacific 3 area in JJA, and (j–l) for the North Pacific 3 area in DJF. Averaged LWP and LTSS are shown in each diagram.

existence of other predominant factors that affect drizzle intensity in the Industrial area during JJA, in addition to COT and CDR. Matsui et al. (2004) reported that not only the amount of aerosol but also the static stability was important for growth from cloud droplets into drizzle.

The vertical inhomogeneity of CDR (larger particles appear in the lower part of clouds) is one possible reason for this observation. Further analyses are required to clarify this issue.

2.3.3 Transition Pattern of Precipitation

Some researchers have considered how the properties of clouds over land and ocean differently affect precipitation efficiency. Leon et al. (2008) analyzed CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) data, and illustrated the global distribution of drizzle frequency as a function of LWP and CDR. N_c was used instead of CDR because this study focused on differences in the amount of aerosol between land (polluted) and ocean (cleaner) regions. Kubar et al. (2009) also investigated the drizzle frequency of water clouds over oceanic areas in the tropics and subtropics as a function of a typical macrophysical variable (LWP) and a typical microphysical variable (N_c). They found that the drizzle frequency increased with LWP when N_c was constant and decreased with increasing N_c and constant LWP. This study focused on the mid-latitudes in the Northern Hemisphere, but more detailed analyses of mid-latitude regions would be valuable.

Figures 2.4 and 2.5 show the $Z_{\max}$ distribution as a function of LWP and N_c , because this study focused on the transition process of drizzle intensity rather than its frequency. Over three ocean regions (Figs. 2.4e–g and 2.5e–g), the drizzle intensity increased with increasing LWP under a constant N_c , and increased with decreasing N_c under a constant LWP. It is important to clarify the physical parameters of clouds to understand the behavior of drizzle over the mid-latitudes as well as over the tropics/subtropics. As the correlation coefficient r_1 between LWP and $Z_{\max}$ (~ 0.6) is greater than r_2 between N_c and $Z_{\max}$ (~ -0.3) in these areas, LWP has a stronger correlation than N_c with drizzle intensity. This correlation is less clear over land areas than over oceanic areas, as shown in Figs. 2.4a–d and 2.5c–d. In particular, high values of $Z_{\max}$ over the Industrial area are scattered during JJA because parameters other than LWP and N_c have strong effects on the drizzle transition process (from cloud droplet to drizzle and precipitation). This is consistent with our hypothesis that there is a more important dominant factor than cloud physical properties, such as COT, CDR, LWP, and N_c over the Industrial area in JJA. The seasonal difference is more obvious over the land areas than over the oceanic areas, with the magnitudes of the correlation coefficients r_1 and r_2 being higher in DJF than in JJA. The land areas in JJA are in the unstable lower LTSS environment, with the exception of Japan. The low specific heat of the land surface would yield unstable conditions due to heating by stronger

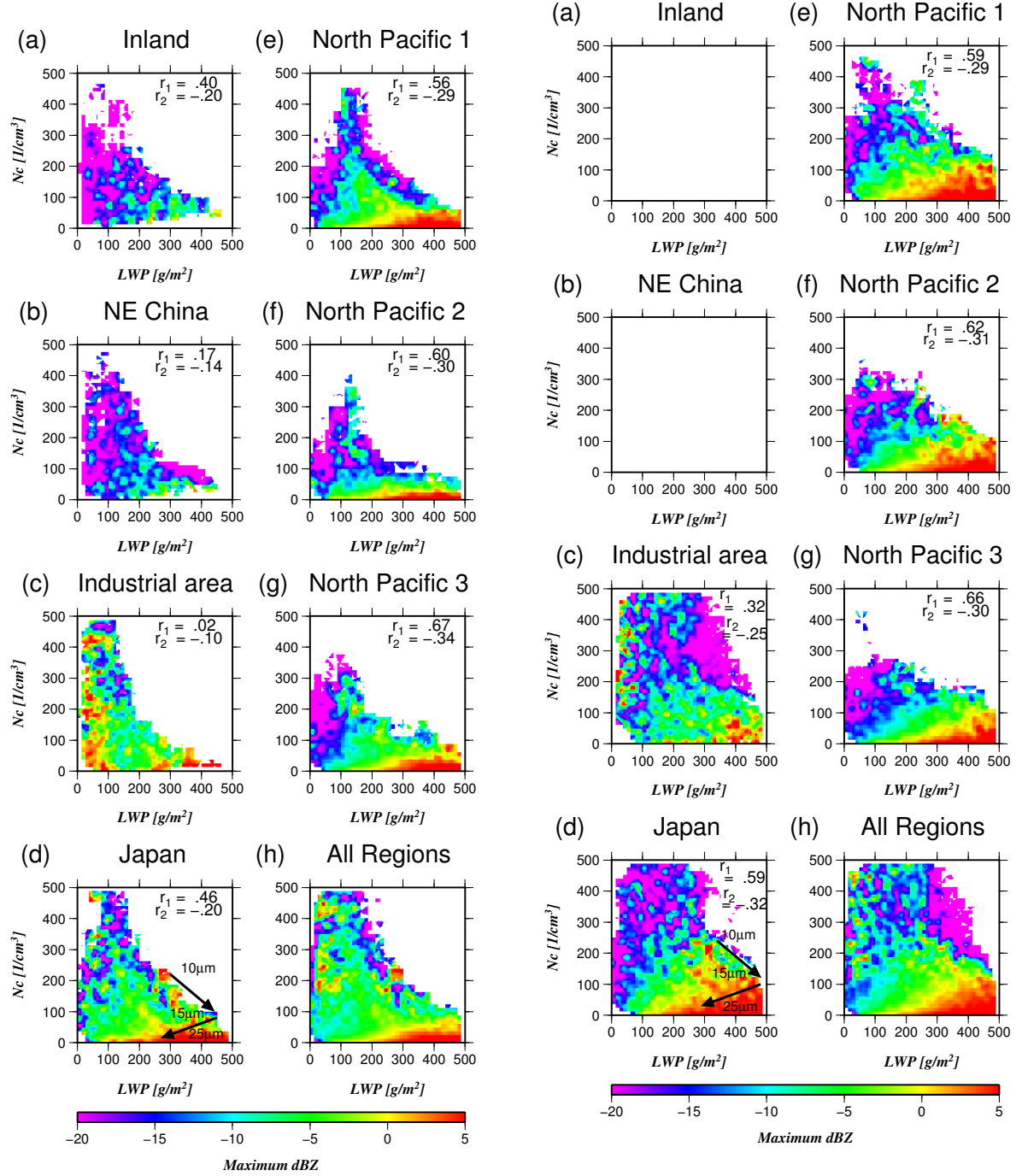


Fig. 2.4: The distribution of maximum radar reflectivity $Z_{\max}$ as a function of LWP and N_c during JJA (a) Inland, (b) NE China, (c) Industrial area, (d) Japan, (e) North Pacific 1, (f) North Pacific 2, (g) North Pacific 3, and (h) the mean value of all regions. r_1 is a correlation coefficient between LWP and $Z_{\max}$ and r_2 is a correlation coefficient between N_c and $Z_{\max}$. The black arrows on (d) indicate one possible interpretation of growing process from cloud droplet to drizzle and raindrop (see text for details).

Fig. 2.5: The distribution of maximum radar reflectivity $Z_{\max}$ as a function of LWP and N_c during DJF (a) Inland, (b) NE China, (c) Industrial area, (d) Japan, (e) North Pacific 1, (f) North Pacific 2, (g) North Pacific 3, and (h) the mean value of all regions. r_1 is a correlation coefficient between LWP and $Z_{\max}$ and r_2 is a correlation coefficient between N_c and $Z_{\max}$. The black arrows on (d) indicate one possible interpretation of growing process from cloud droplet to drizzle and raindrop (see text for details). The diagrams of (a) Inland and (b) NE China are not shown because no data are available.

shortwave radiation in the JJA season. Such local heating may result in forced precipitation. This is responsible for the scattered distribution of high $Z_{\max}$ values. In addition, variations in the dynamics over land areas (e.g., vertical velocity) would also be associated with this seasonal difference.

Values of $Z_{\max}$ greater than 0 dBZ_e (orange and red in Fig. 2.4) are uncommon in the Inland and NE China areas during JJA, which indicates very few precipitating clouds. Over these regions in DJF, generally only few water clouds are observed due to low temperature and/or low water vapor levels. In the Industrial area, there are some occasions when N_c is larger than 500 cm⁻³, and $Z_{\max}$ values are lower as N_c becomes larger during DJF. Even LWP values, which are more strongly correlated with drizzle intensity, are larger. These findings suggest that the cloud lifetime increases due to storage of water within the cloud layer. These findings are also observed in Japan (Fig. 2.5d), where a significant transition pattern appear as follows: LWP of 300 g m⁻² and N_c of 250 cm⁻³, to LWP of 450 g m⁻² and N_c of 100 cm⁻³, to LWP of 300 g m⁻² and N_c of 15 cm⁻³, as shown by the black arrows in Figs. 2.4d and 2.5d. LWP values increase to 400–500 g m⁻² as N_c values decrease because drizzle occurs only inside the cloud layer with no loss of water. At the same time, CDR values increase slowly within the range of 10–15 μm and then rapidly to larger values (15–25 μm), which leads to precipitation. The conditions in Japan are not as clean as in the three oceanic regions, but are not as polluted as in the Industrial area, which is likely the reason for this V-shaped transition pattern.

2.3.4 Cloud Vertical Structure

Cloud geometrical thickness is a cloud macrophysical variable, in addition to the cloud-top height and LWP. Over the tropical ocean, cloud-top height is offset by a constant from the cloud geometric thickness, because the cloud base height is almost constant (e.g., Kubar et al., 2009). Cloud base height is, however, not always constant over mid-latitudes, particularly over the land. Therefore, this study use cloud geometrical thickness as a representative macrophysical variable. In fact, cloud geometrical thickness has a robust correlation with $Z_{\max}$ (0.28–0.83; shown in Fig. 2.6), which is an index of precipitation intensity, stronger than the relationship between cloud-top height and $Z_{\max}$ (0.04–0.63). However, it should be noted that the “cloud geometrical thickness” mentioned here does not always accurately represent the cloud thickness. Specifically, in some cases of non-precipitating cloud, determination of the cloud base is difficult because the reflectivity at this point is too weak to be observed. However, in the case of precipitating cloud,

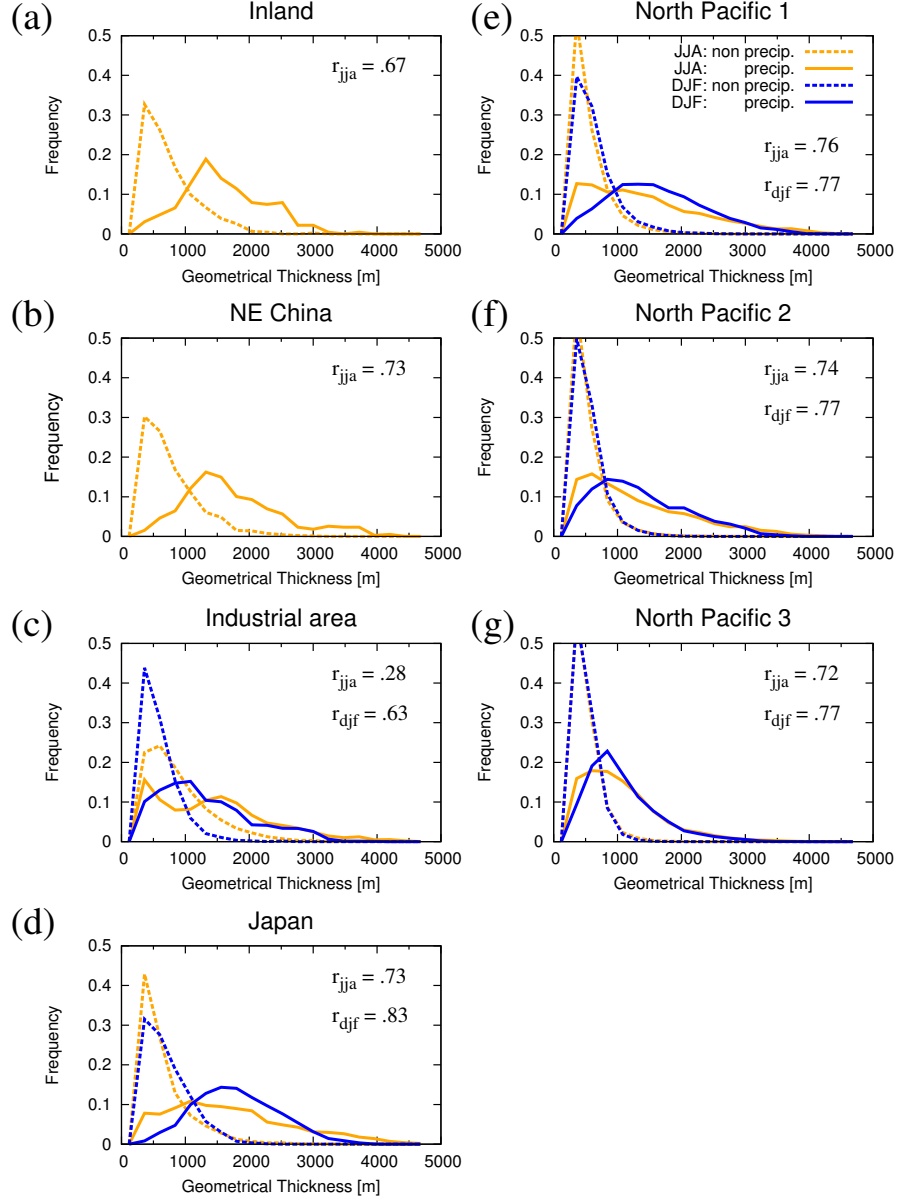


Fig. 2.6: PDFs of cloud geometrical thickness for non-precipitating cloud (dotted line) and drizzling/precipitating cloud (solid line). r_{jja} is the correlation coefficient between cloud geometrical thickness and $Z_{\max}$ in JJA season, and r_{djf} is that in the DJF season. The DJF values of (a) Inland and (b) NE China are not shown because no data are available.

the detected value would include not only the cloud but also some of the precipitating layer. Thus, the “cloud geometrical thickness” represents the detected hydrometer thickness.

The PDFs of cloud geometrical thickness are shown in Fig. 2.6. Solid and dotted lines represent drizzling/precipitating and non-precipitating cloud, respectively. The correlations between cloud geometrical thickness and $Z_{\max}$ for JJA and DJF are denoted as r_{jja} and r_{djf} , respectively. Almost all of the non-precipitating clouds have geometrical thickness less than 1000 m, and the

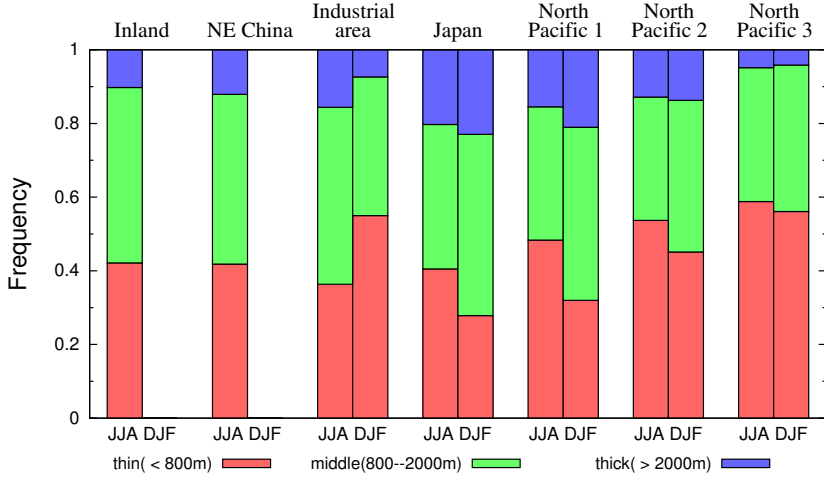


Fig. 2.7: Histogram of cloud geometrical thickness. Thin (red), middle (green), and thick (blue) clouds are defined using threshold values of 800 and 2000 m. The DJF values of (a) Inland and (b) NE China are not shown because there is no available data.

clouds with precipitation are ~ 500 – 1000 m thicker. This trend and strong correlation between cloud geometrical thickness and $Z_{\max}$ suggest the importance of cloud geometrical thickness for the occurrence of precipitation. The modal cloud geometrical thickness of the non-precipitating category is ~ 500 m for all seven regions during both seasons. On the other hand, the precipitating clouds have large seasonal variability. For example, oceanic clouds (Fig. 2.6e–g) become thicker in DJF. Figure 2.7 shows a histogram of cloud geometrical thickness for thin (< 800 m; red), middle (800–2000 m; green), and thick (≥ 2000 m; blue) clouds, which correspond roughly to non-precipitating, drizzling, and precipitating clouds, respectively. The LTSS values listed in Table 2.1, which represent the air stability, tend to be consistent with the cloud geometrical thickness. More specifically, middle or thicker clouds exist predominantly in the unstable environment over the Industrial area in JJA (i.e., $LTSS = 12.2$ K). Conversely, in the stable environment in DJF (i.e., $LTSS = 19.6$ K), thinner clouds are predominant. Similar to this tendency, the cloud geometrical thickness, which reflects the seasonal difference in LTSS, is also seen among other regions.

Lebsock et al. (2008) confirmed that high-aerosol conditions tend to decrease LWP in non-precipitating clouds, and the magnitude of the reduction in LWP is greater under the unstable low LTSS environment. These findings suggest the importance of LWP and thermodynamics to understanding aerosol–cloud interactions (L’Ecuyer et al., 2009). The present study further

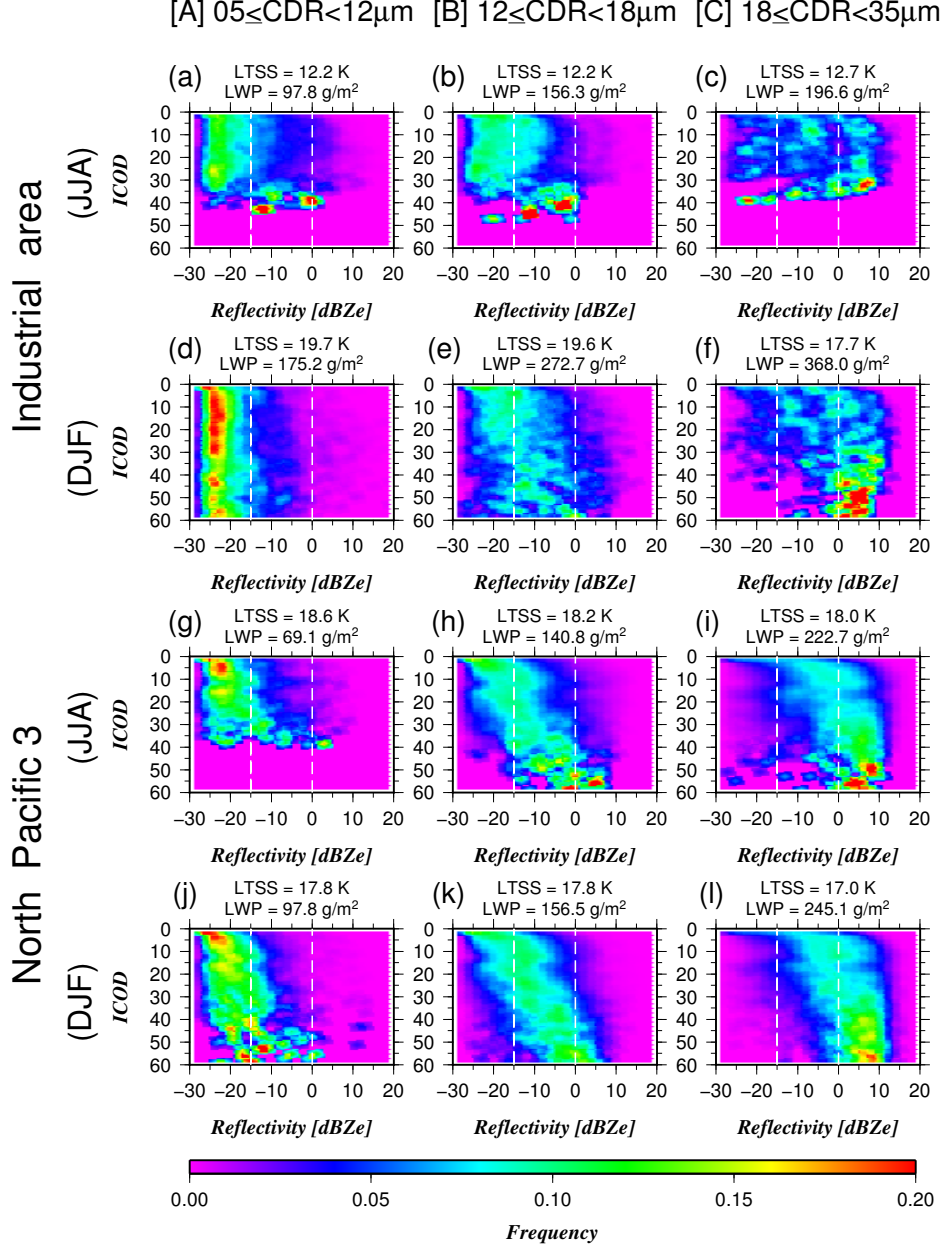


Fig. 2.8: Contoured frequency by optical-depth diagrams (CFODDs) as a function of CDR, [A] $5\text{--}12 \mu\text{m}$, [B] $12\text{--}18 \mu\text{m}$, [C] $18\text{--}35 \mu\text{m}$. (a–c) are for the Industrial area in JJA, (d–f) for the Industrial area in DJF, (g–i) for the North Pacific 3 area in JJA, and (j–l) for the North Pacific 3 area in DJF. Two white dotted lines are drawn as threshold radar reflectivity values, -15 dBZ_e and 0 dBZ_e , which are taken as the boundaries between cloud particles and drizzle, and between drizzle and rain, respectively. Averaged LWP and LTSS are also shown in each CFODD.

investigated the cloud vertical structure based on a comparison with the atmospheric conditions (pristine or polluted) associated with LWP and LTSS. Use of the CFODD to illustrate cloud vertical structure facilitates identification of associations with cloud optical properties, particularly

for single-layered water clouds (e.g., Nakajima et al., 2010; Suzuki et al., 2010). In general, the vertical and horizontal axes are allocated to geometrical height and radar reflectivity, respectively, when illustrating the frequency of the vertical radar profile. CFODD visualization methods apply the in-cloud optical depth (ICOD) as the vertical axis instead of altitude. In this way, normalization of the vertical coordinate by ICOD facilitates interpretation focusing on optical properties using composited clouds of different geometrical thicknesses. The data of layered optical depth was obtained from the 2B-TAU product.

CFODDs of each CDR bin ([A] 5–12 μm , [B] 12–18 μm , [C] 18–35 μm) over the Industrial area and North Pacific 3 are presented in Fig. 2.8. Although LTSS is correlated with cloud geometrical thickness, as mentioned earlier, LTSS seems to have little relation with the cloud growth process, because the values are almost identical among the three CDR bins. The CFODDs show that the LWP increases monotonically with increasing CDR, which corresponds to the transition from cloud particle (category [A]) to drizzle (category [B]), and raindrop (category [C]). That is, CDR bin [A] represents evaporation and condensation processes, and CDR bins [B] and [C] represent mainly collision and coalescence processes. Therefore, an increase in LWP with an increase in CDR is expected. However, the rate of increase of LWP differs significantly between the Industrial area and North Pacific 3, as shown in Table 2.2. That is, the rate of increase over North Pacific 3 is greater than that over the Industrial area. This result implies that the clouds over North Pacific 3 are more efficient than those over the Industrial area in terms of cloud droplet growth. Over the Industrial area in DJF, which is in the stable and high LTSS environment, non-precipitating clouds are dominant (61.5 %; see Table 2.1) and contain much cloud water, as depicted in Fig. 2.8d. This may suggest the occurrence of the second indirect effect (Albrecht, 1989). Under such high-LWP and small-CDR conditions, cloud albedo can also

Table 2.2: LWP and its rate of increase for each CFODD.

	[A] $05 \leq \text{CDR} < 12 \mu\text{m}$	[B] $12 \leq \text{CDR} < 18 \mu\text{m}$	[C] $18 \leq \text{CDR} < 35 \mu\text{m}$
Industrial area			
JJA LWP [g m^{-2}]	97.8	156.3	196.6
Rate of increase	1.60	1.26	
DJF LWP [g m^{-2}]	175.2	272.7	368.0
Rate of increase	1.56	1.35	
North Pacific 3			
JJA LWP [g m^{-2}]	69.1	140.8	222.7
Rate of increase	2.04	1.58	
DJF LWP [g m^{-2}]	97.8	156.5	245.1
Rate of increase	1.60	1.57	

increase, as can be seen in the following Eq. (2.3), which is another form of Eq. (2.1),

$$\tau_c = \frac{9\text{LWP}}{5r_e}. \quad (2.3)$$

In fact, the COT in DJF ($\tau_c = 35.9$) is much higher than that in JJA ($\tau_c = 19.5$).

Non-precipitating clouds were observed mainly in the smallest CDR bin ($\text{CDR} < 12 \mu\text{m}$), and an obvious transition of the CFODD to drizzle ($12 \mu\text{m} \leq \text{CDR} < 18 \mu\text{m}$) and rain ($18 \mu\text{m} \leq \text{CDR}$) phases. In addition, there is a clear difference between the CFODDs of the Industrial area and North Pacific 3, with regard to the transition process for drizzling clouds. More specifically, the CFODDs over the polluted land area transit ICOD mainly from near the cloud-top to the cloud-base, while those over the ocean transit mainly in the deeper ICOD region (approximately over 30). This feature is consistent with some previous reports (e.g., Nakajima et al., 2010; Suzuki et al., 2010, 2011). This characteristic of CFODDs can be interpreted as a result of suppression of precipitation due to high concentrated aerosols around the cloud base (large part of ICOD) over the Industrial area. DJF is a dry season over mid-latitudes in the Northern Hemisphere, and the stable and high LTSS environment results in a high aerosol concentration near the surface. Therefore, an aerosol–cloud interaction may occur that results in weaker radar reflectivity in the larger ICOD region. This may be one of the hypotheses, and further analysis (e.g., sensitivity experiments using numerical modeling) is required in order to enhance the credibility. It is also possible that the difference in cloud vertical structure between land and ocean is caused by the difference in updraft strength (Nakajima et al., 2010), or other meteorological factors as well. The mission of “Earth Clouds, Aerosols and Radiation Explorer (Earth-CARE),” which will start in 2016, is helpful because it will equip the CPR with Doppler speed sensor functions (e.g., Sy et al., 2013; Nakatsuka et al., 2012; Schutgens, 2008) that can detect vertical velocity. In addition, numerical modeling experiments are required for further understanding of the aerosol–cloud–radiation interaction.

Lebsock et al. (2008) emphasized the importance of performing investigations on regional and seasonal scales in both numerical modeling and observational studies to gain a more detailed understanding of cloud dynamics. Suzuki et al. (2013b) also suggested that the complex behavior of CFODDs at different latitudes (see their Fig. S3) and models could not reproduce the satellite-observed CFODDs due to a lack of knowledge of parameterization of cloud dynamics at different latitudes. The results of the present study based on regional and seasonal analysis associated with the aerosol–cloud interaction will contribute to the improvement of cloud physical parameterization in numerical models.

The effects of spatial difference of meteorology on aerosol–cloud interaction were not considered in our study; therefore, further analyses are necessary. We must be careful about following two ideas: one is the fact that genuine aerosol–cloud interactions may behave differently under different meteorological conditions; and the other is the fact that meteorology may drive aerosol–cloud relationships (even in the absence of any aerosol–cloud interactions). Such the meteorological gradients sometimes cause spurious correlations (Grandey and Stier, 2010).

For example, the difference in the autoconversion rate over land and ocean, or in JJA and DJF, may provide some insight into the indirect aerosols effects (e.g., Stephens and Haynes, 2007; Sorooshian et al., 2013). Although the data presented here are insufficient to link the ocean versus land differences to aerosol effects, further studies to determine the effects of atmospheric conditions (i.e., aerosol concentration, static stability) on cloud physical structure would be valuable.

2.4 Summary Perspectives

This chapter conducted a comparative study of the physical properties of water clouds over the region from East Asia to the North Pacific in the mid-latitudes based on CloudSat/CPR and Aqua/MODIS retrievals. In addition to confirming several known characteristics regarding cloud physical properties, such as larger N_c , smaller CDR, and larger COT values over land, it was found that the cloud differences over land versus the ocean are more obvious during DJF than JJA.

In the pristine area, a clear tendency for lower to higher COT–CDR with rising precipitation categories was found during both JJA and DJF. However, this transition pattern does not appear clearly in the polluted area during JJA, and precipitation occurs even in the lower COT–CDR region.

An investigation of the transition process of precipitation reveals that during DJF the polluted areas have larger N_c values, and the clouds could contain much more LWP with larger N_c values than during JJA. Oceanic cloud properties over the mid-latitudes do not change significantly between the two seasons, and their behavior is similar to that of oceanic clouds over the tropics/subtropics. However, considerable seasonal differences was observed over land.

Such differences also appear in the LTSS. Although the LTSS is correlated with cloud geometrical thickness, it is less important for the cloud growth process. On the other hand, LWP increases monotonically with growing CDR. However, a smaller rate of increase in LWP was

confirmed over polluted land. In addition, a difference in “contoured frequency by optical-depth diagram” (CFODD) between the pristine oceanic area and the polluted land area was found, implying the aerosol–cloud interaction. However, this study cannot completely exclude the possibility that other meteorological factors may be responsible for the differences between land and ocean.

To clarify these differences in cloud properties and drizzle characteristics between land and ocean, and between the tropics/subtropics and mid-latitudes, it is important to estimate the radiation budget accurately. Some of the characteristics of aerosol–cloud interaction were determined based only on satellite data. However, composite studies with numerical modeling (e.g., sensitivity experiments for the influence of aerosol and atmospheric stability to cloud physics) are required to gain a detailed understanding of the aerosol–cloud interaction. This study does not preclude the possible effect of spatial gradient changes in the meteorology on aerosol–cloud interaction, and further analyses taking such environmental conditions into consideration are required.

3 Sensitivity Experiments on Cloud-to-Rain Conversion Process with Global Aerosol-Climate Model

Abstract

We examined the performance of autoconversion (mass transfer from cloud water to rainwater by the coalescence of cloud droplets) schemes in warm rain, which are commonly used in GCMs. To exclude biases in the different treatment of the aerosol–cloud–precipitation–radiation interaction other than that of the autoconversion process, sensitivity experiments were conducted within a single-model framework using an aerosol climate model, MIROC-SPRINTARS. The liquid water path (LWP) and cloud optical thickness have a particularly high sensitivity to the autoconversion schemes, and their sensitivity is of the same magnitude as model biases. In addition, the ratio of accretion (growth of raindrops via collision/coalescence with cloud droplets) to autoconversion (Acc/Aut ratio), a key parameter in the examination of the balance of microphysical conversion processes, also has a high sensitivity globally depending on the scheme used. Although the Acc/Aut ratio monotonically increases with increasing LWP, significantly lower ratio is observed in Kessler-type schemes. Compared to satellite observations, a poor representation of cloud macrophysical structure and optically thicker low cloud are found in simulations with any autoconversion scheme. As a result of the cloud–radiation interaction, the difference in the global mean net cloud radiative forcing (NetCRF) among the schemes reaches 10 W m^{-2} . The discrepancy between the observed and simulated NetCRF is especially large with a high LWP. The potential uncertainty in the parameterization of the autoconversion process is non-negligible, but none of the tested schemes significantly improve the bias in the cloud radiative effect yet. This means that more fundamental errors are still left in other processes of the model.

Keywords: autoconversion, accretion, microphysics, cloud–radiation feedback

Key Points:

- Commonly used autoconversion schemes are evaluated in a single model framework
- The ratio of accretion to autoconversion is sensitive to the scheme used

- There are more fundamental errors in other processes of the model in addition to the microphysical representation
-

3.1 Introduction

The fact that aerosol particles directly/indirectly affect the Earth’s radiation budget and cloud microphysics is widely known as the aerosol–radiation interaction (e.g., Coakley et al., 1987; Charlson et al., 1992) and aerosol–cloud interaction (e.g., Twomey, 1977; Albrecht, 1989). The global distribution of clouds is one of the most important factors determining the Earth’s radiation budget. For example, stratocumulus in the marine boundary layer (MBL) cools the climate system (Hartmann et al., 1992). However, because the grid intervals of general circulation models (GCMs) are not fine enough to reproduce the spatial distribution of each cloud due to the constraints imposed by computational resource limits, GCMs depend strongly on parameterizations. Therefore, most GCMs cannot reproduce low stratocumulus very well, which is a critical issue for GCMs and needs to be improved (Medeiros and Stevens, 2011; Nam et al., 2012).

The transition process for warm rain (i.e., from cloud water to rainwater) in model simulations is the so-called autoconversion, which is a strongly nonlinear process. The autoconversion parameterization is given below (e.g., Beheng, 1994, hereafter Be94), as a function of liquid water content (L_c : kg m^{-3}) and cloud droplet number concentration (N_c : m^{-3}):

$$\left(\frac{\partial L_r}{\partial t}\right)_{aut} = C_{aut} L_c^\alpha N_c^\beta \quad (3.1)$$

where L_r is the rainwater content (kg m^{-3}), and C_{aut} , α , and β are constants. The growth via collision/coalescence of rain drops with cloud droplets is so-called accretion. The accretion parameterization is given below (e.g., Khairoutdinov and Kogan, 2000, hereafter KK00) as a function of L_c and L_r :

$$\left(\frac{\partial L_r}{\partial t}\right)_{acc} = C_{acc} (L_c L_r)^\gamma \quad (3.2)$$

where C_{acc} and γ are constants. The nonlinearity of the autoconversion process is greater than that of accretion (i.e., $\alpha, \beta > \gamma$) in general. Equations (3.1) and (3.2) describe the rate of mass transfer from cloud water to rainwater by the collision/coalescence of cloud droplets (autoconversion) and the collision of cloud droplets by raindrops (accretion), respectively. Consequently, the total conversion rate can be written as a summation of the autoconversion and accretion terms:

$$\left(\frac{\partial L_r}{\partial t}\right)_{conv} = \left(\frac{\partial L_r}{\partial t}\right)_{aut} + \left(\frac{\partial L_r}{\partial t}\right)_{acc}$$

Because the strong nonlinearity of the autoconversion scheme results in a high susceptibility to subgrid-scale variability (Morrison and Gettelman, 2008; Lebsock et al., 2013), the parameterizations with regard to the conversion processes are vitally important for GCMs (Weber and Quaas, 2012).

This parameterization and coarse resolution leads to uncertainty in simulations. The magnitude of the radiative forcing by the aerosol–cloud–precipitation–climate interaction still has major uncertainty, as summarized in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC 2013). The magnitude of the dependence of the autoconversion rate on L_c and N_c (i.e., α and β) is involved in this uncertainty (e.g., Menon et al., 2003; Lohmann and Feichter, 2005), because the effects of aerosols on cloud lifetime and precipitation susceptibility (e.g., Wood et al., 2009; Sorooshian et al., 2009; Jiang et al., 2010) are strongly controlled by the autoconversion scheme as described by equation (3.1). For example, Penner et al. (2006) and Quaas et al. (2009) conducted a GCM intercomparison to identify the mechanisms of uncertainty in the aerosol–cloud–radiation interaction among the models. They discussed the issues associated with the different treatments of the autoconversion schemes, and quantitatively presented the uncertainty. Franklin (2008) demonstrated that the frequency of non-precipitation/drizzle/precipitation was sensitive to the autoconversion scheme. Suzuki et al. (2011, 2013a) reported that models generally tend to predict drizzle/rain with a high frequency compared to A-Train satellite observations, and suggested that microphysical vertical structure is the key to improve the model performance. Nam et al. (2012) and Nam and Quaas (2013) also reported an inconsistency in the radar reflectivity profile (i.e., microphysical vertical structure) between observations and GCM simulations. The development of the parameterization for the aerosol–cloud–precipitation–climate interaction is still a challenging topic, and an appropriate evaluation of the microphysical schemes is required.

Wood (2005a) estimated autoconversion and accretion rates for MBL clouds from the integration of the stochastic collection equation (SCE) with the observed drizzle size distribution by aircraft measurements, and examined the differences among several parameterizations of autoconversion and accretion. The schemes proposed by Tripoli and Cotton (1980) (hereafter TC80) overestimated the autoconversion rate, whereas a good performance was achieved by KK00 scheme. Gettelman et al. (2013) demonstrated that GCMs overestimate the autoconversion rate relative to the observation that estimated from Variability of the American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study (VOCALS) aircraft measurements (Wood et al., 2011;

Terai and Wood, 2013). On the other hand, accretion schemes generally performed better than autoconversion. Chuang et al. (2012) examined the performance of autoconversion and nucleation parameterizations using a GCM. They found that the treatment of autoconversion was more sensitive than that of nucleation. In addition, they discussed microphysical feedbacks and calculated the indirect effects of sulfur aerosols on the diurnal scale simulation (~ 3 days). Although their results of a short-term simulation, which aimed to preclude compensation of multiple feedback mechanisms, provided valuable information for model improvement, the influence of differences in microphysical parameterizations on climate change has not yet been investigated.

Therefore, we evaluated the autoconversion schemes and their potential errors in microphysical processes using several satellite datasets, which will provide a helpful key to improve cloud parameterization. It should be noted that autoconversion and accretion process rates cannot be observed directly from current satellite remote sensing, and the global statistics of them are left implicit. Because an intercomparison of GCMs cannot completely preclude biases from different treatments of aerosol–cloud–precipitation interaction in each model, we conducted an intercomparison of autoconversion schemes only within the framework of one specific model (i.e., not a model intercomparison), and then considered the effects of cloud micro- and macrophysics and the global radiation budget on climatology.

The remainder of this paper is organized as follows. Section 3.2 briefly describes the model and the autoconversion schemes examined in the study. In Section 3.3, the results of the global and regional analyses from sensitivity experiments are presented from the perspective of process rate, cloud properties, and radiation budget. Finally, the findings of the study and a discussion are given in Section 3.4.

3.2 Model and Experiments

3.2.1 General Model Description

A global aerosol–climate model, Spectral Radiation-Transport Model for Aerosol Species (SPRINT-ARS), which can reproduce the global distribution of the main aerosol species in the troposphere (Takemura et al., 2000, 2002, 2005, 2009) was used. SPRINTARS has been coupled with a comprehensive atmospheric GCM (AGCM), “Model for Interdisciplinary Research On Climate (MIROC 5.2) (Watanabe et al., 2010, 2011)” (MIROC-SPRINTARS).

SPRINTARS explicitly predicts the mass mixing ratios of carbonaceous particles (black carbon (BC) and organic matter (OM)), sea salt, soil dust, and sulfate. In addition, the precursor

gases of sulfate (i.e., sulfur dioxide (SO_2) and dimethyl sulfide (DMS)) are also considered tracers. The aerosol transport processes consist of emission, advection, diffusion, chemical reaction of sulfate, dry and wet deposition, and gravitational settling. The sources of BC and OM are considered to be biomass burning, consumption of fossil fuel and fuel wood, agricultural activities, and gas-to-particle conversion of terpene (Takemura et al., 2000). The size of sea salt aerosols is treated from $0.1\ \mu\text{m}$ to $10\ \mu\text{m}$ according to a lognormal size distribution with four effective radii (i.e., 0.178 , 0.562 , 1.780 , and $5.620\ \mu\text{m}$), and their generation is predicted as a function of wind velocity at 10 m height (V_{10} ; for details see *Appendix B* of Takemura et al. (2009)). The emission of soil dust is controlled by the vegetation type based on International Satellite Land Surface Climatology Project Initiative II (ISLSCP II) C4 vegetation data (Still, 2003), and depends on V_{10} , soil moisture, leaf area index (LAI), and snow accumulation (for details see *Appendix A* of Takemura et al. (2009)). The sources of sulfur species are fossil fuel consumption, volcanic activity, activity of oceanic phytoplankton, and biomass burning.

MIROC-SPRINTARS applies the flux-form semi-Lagrangian scheme (Lin and Rood, 1996; Ritchie, 1986) for the tracer transport process. In addition, the transport is also included in the Arakawa-Schubert parameterization (Arakawa and Schubert, 1974), which simulates convective clouds. In chemical reaction processes, monthly mean global concentrations of hydroxyl radical (OH), ozone (O_3), and hydrogen peroxide (H_2O_2) calculated by a global chemical climate model (CHASER) (Sudo et al., 2002) are used.

MIROC-SPRINTARS uses minimal advanced treatments of surface interaction and runoff (MATSIRO) for the land component (Takata et al., 2003). MATSIRO predicts the interchange of heat and water between land and atmosphere, which changes the land surface albedo and global radiation budget. The effects of surface water melting on variation in ice sheet albedo (e.g., Bougamont et al., 2005) is considered in the model, by assuming that the albedo is a function of the water content above the ice.

The radiative transfer in MIROC-SPRINTARS is calculated by mstrnX (Sekiguchi and Nakajima, 2008), which can consider the effects of cloud overlap (e.g., Morcrette and Fouquart, 1986). The aerosol-radiation interaction is evaluated based on the mass mixing ratios of aerosols with prescribed size distributions, hygroscopic growth of aerosol particles, mode radii depending on the relative humidity, refractive indices, extinction and absorption coefficients, and the other radiative factors for each aerosol component (Takemura et al., 2005). N_c is calculated based on the Köhler theory (Takemura et al., 2005), which depends on the aerosol number concentration,

updraft velocity, size distributions, and saturation condition of water vapor. The aerosol–cloud–precipitation interaction is also evaluated in the autoconversion scheme proposed by Berry (1968) (hereafter Be68). The details of the cloud microphysics and autoconversion process are described in the following subsections (cf. Section 3.2.2 and 3.2.3).

3.2.2 Cloud Microphysics

The framework of microphysical parameterization in MIROC-SPRINTARS is composed of a prognostic large scale condensation (LSC) scheme (Watanabe et al., 2009), which is coupled with the bulk microphysical scheme proposed by Wilson and Ballard (1999). This scheme applies a probability distribution function (PDF) of temperature and total water content, hence the subgrid-scale inhomogeneity of cloud can be reproduced. The distribution of cloud and precipitation has been improved using the PDF scheme against the previous version (MIROC3.2), which applied a diagnostic LSC scheme based on Le Treut and Li (1991).

MIROC-SPRINTARS microphysics is based on the Wilson and Ballard (1999) scheme, which considers four water quantities, i.e., vapor, liquid cloud droplets, raindrops, and frozen water. The transfer processes are as follows: condensation/evaporation (between cloud liquid water and vapor); homogenous and heterogenous nucleation (from vapor to ice crystals); deposition and sublimation (from ice to vapor); riming, i.e., the collection of cloud liquid water by ice (from liquid to ice); capture of raindrops by ice particles (from rain to ice); melting of snow to rain (from ice to rain); evaporation of rain (from rain to vapor); and warm rain microphysics by autoconversion and accretion (from liquid water to rainwater). The current version of MIROC-SPRINTARS does not consider drizzle and rainwater as prognostic variables. The predicted rainwater instantaneously falls out to the surface as precipitation within a single time step (Ghan and Easter, 1992).

MIROC-SPRINTARS explicitly predicts the number concentrations for cloud droplet and ice by the prognostic equations (*Appendix C* of Takemura et al. (2009)). Both cloud droplets and ice crystals are treated as two-moment schemes by predicting their mass mixing ratios and number concentrations. Those for both liquid and ice clouds vary via the processes of nucleation, accretion, riming, and aggregation processes. The nucleation of cloud droplets is based on the parameterization by Abdul-Razzak and Ghan (2000), which depends on the aerosol particle number concentrations, size distributions, curvature effect, solute effect of each aerosol component, and the updraft velocity. The further description of parameterizations of the aerosol–cloud inter-

Table 3.1: List of autoconversion schemes examined in this study. Units are kg m^{-3} for liquid water content L_c and rainwater content L_r , and kg kg^{-1} for the cloud water mixing ratio q_c . The units of air density ρ_a and liquid water density ρ_w are kg m^{-3} .

Schemes	Autoconversion rate ($\text{kg m}^{-3} \text{s}^{-1}$)	Models
Berry (1968)	$\frac{3.5 \times 10^{-2} L_c^2}{0.12 + 1.0 \times 10^{-12} \frac{N_c}{L_c}}$	MIROC5; NICAM-SPRINTARS
Tripoli and Cotton (1980)	$\frac{0.104 g E_{cr}}{\mu \rho_w^{1/3}} L_c^{7/3} N_c^{-1/3} H(q_c - q_{crit})$	HadGEM2; GFDL CM3
Beheng (1994)	$6.0 \times 10^{28} n^{-1.7} (L_c \times 10^{-3})^{4.7} (N_c \times 10^{-6})^{-3.3}$	ECHAM5
Khairoutdinov and Kogan (2000)	$1350 L_c^{2.47} (N_c \times 10^{-6})^{-1.79} \rho_a^{-1.47}$	ECHAM5-HAM; PNNL MMF
Liu and Daum (2004)	$\left(\frac{3}{4\pi\rho_w}\right)^2 \kappa_2 \beta_6^6 L_c^3 N_c^{-1} H(R_6 - R_{6c})$	CSIRO Mark3

action on water and ice clouds is summarized in Takemura et al. (2005, 2009). A more detailed description of the model and its cloud microphysics are also presented in Watanabe et al. (2009, 2010).

3.2.3 Experimental Design

Table 3.1 summarizes the autoconversion schemes examined in this study, which are commonly incorporated into GCMs. Be68 and Be94 indicated the relationship between L_c and N_c in the autoconversion rate based on the SCE, and KK00 derived it from large-eddy simulation (LES). MIROC-SPRINTARS uses the Be68 scheme for cloud autoconversion as a default. The Non-hydrostatic Icosahedral Atmospheric Model (NICAM), which is a global cloud resolving model developed in Japan (e.g., Tomita and Satoh, 2004; Satoh et al., 2008), also uses this scheme. The KK00 scheme is also widely used, for example, in ECHAM5-HAM (e.g., Stier et al., 2005; Croft et al., 2009; Zhang et al., 2012a) and PNNL-MMF (e.g., Wang et al., 2011a,b).

Kessler suggested a simplified parameterization for the conversion from cloud water to rainwater when the cloud water mixing ratio q_c exceeds a threshold value (q_{crit}). A Kessler-type parameterization was used by many researchers (e.g., Manton and Cotton, 1977; Tripoli and Cotton, 1980; Liou and Ou, 1989; Baker, 1993; Boucher et al., 1995) for the further improvement of cloud microphysical processes. For examples, the TC80 scheme considered the parameters of collection efficiency (E_{cr} : assumed value of 0.55), acceleration of gravitation (g), and dynamic viscosity (μ) to derive the autoconversion rate. This scheme is applied, for example, in HadGEM2 (e.g., Collins et al., 2011) and GFDL CM3 (e.g., Donner et al., 2011). Liu and Daum (2004)

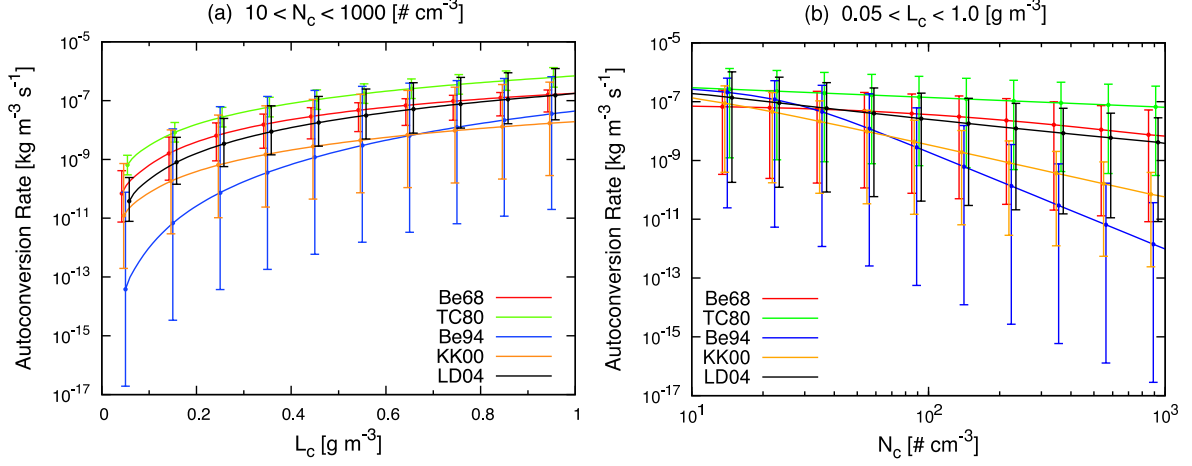


Fig. 3.1: autoconversion rate ($\text{kg m}^{-3} \text{s}^{-1}$) as a function of the (a) liquid water content L_c , and (b) cloud droplet number concentration N_c . The curves represent (a) $N_c = 100$ (cm^{-3}) and (b) $L_c = 0.1$ (g m^{-3}). The error bar represents the range of variation with (a) N_c from 10 to 1000 (cm^{-3}) and (b) q_c from 0.05 to 1.0 (g m^{-3}).

(hereafter LD04) proposed a new Kessler-type parameterization for applying the mean radius of the sixth moment (R_6) as a threshold parameter (R_{6c}). Further discussion of the Kessler-type threshold function is provided elsewhere (e.g., Wood and Blossey, 2005; Liu et al., 2006). This scheme was examined by Rotstayn and Liu (2005) in CSIRO Mark3.

The variation in the autoconversion rate (unit: $\text{kg m}^{-3} \text{s}^{-1}$) as a function of L_c and N_c in each scheme is shown in Fig. 3.1. The curves represent the autoconversion rates for typical values in MBL clouds (i.e., $N_c = 100 \text{ cm}^{-3}$, $L_c = 0.5 \text{ g m}^{-3}$). The variation in N_c from 10 to 1000 cm^{-3} (Fig. 3.1a) and L_c from 0.05 to 1.0 g m^{-3} (Fig. 3.1b) is shown with error bars, which indicate the magnitude of nonlinear dependency on each variable. The schemes with a lower nonlinear dependency on N_c such as Be68 ($\beta = -1.0$), TC80 ($\beta = -0.33$), and LD04 ($\beta = -1.0$) have narrower bars than the stronger ones such as Be94 ($\beta = -3.3$) and KK00 ($\beta = -1.79$). Regarding the parameter α in L_c , Be94, which has the strongest nonlinearity ($\alpha = 4.7$), has the widest bar.

This study examined the performance of the five schemes summarized above based on a MIROC-SPRINTARS simulation. The horizontal and vertical resolutions are T42 (approximately $2.8^\circ \times 2.8^\circ$ in latitude and longitude), and 20 layers (σ -pressure coordinate system: 0.995, 0.980, 0.950, 0.900, 0.830, 0.745, 0.650, 0.549, 0.454, 0.369, 0.295, 0.230, 0.175, 0.124, 0.085, 0.060, 0.045, 0.035, 0.025, and 0.008), respectively. The standard time step is 20 min for each time integration, which becomes shorter in cases with an unstable CFL condition. All sensitivity experiments were integrated for 6 years and analyzed for the last 5 years as a climatology.

Table 3.2: Simulated global mean climatology of physical parameters from the different autoconversion schemes. Observational data are obtained from GPCP version 2.2 (for the period 2001–2013) for precipitation flux, ISCCP D2 (for the period 2005–2009) for cloud cover, Aqua-MODIS Level 3 (for the period 2006–2010) for cloud optical thickness (COT), cloud effective radius (CER), liquid water path (LWP), and aerosol optical thickness (AOT), NOAA-18 interpolated OLR (for the period 2005–2009), and CERES EBAF–TOA Ed2.8 (for the period 2001–2013) for shortwave/longwave/net cloud radiative forcing (SWCRF, LWCRF, NetCRF). N_c burden was estimated from AVHRR observation by Han et al. (1998).

	Model-based schemes			Kessler-type schemes		Obs.
	Be68	Be94	KK00	TC80	LD04	
Autoconversion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	6.9×10^{-6}	6.4×10^{-6}	6.2×10^{-6}	8.8×10^{-6}	6.5×10^{-6}	–
Accretion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	2.1×10^{-6}	2.2×10^{-6}	2.4×10^{-6}	1.1×10^{-6}	1.9×10^{-6}	–
Total conversion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	9.0×10^{-6}	8.6×10^{-6}	8.6×10^{-6}	9.9×10^{-6}	8.4×10^{-6}	–
Acc/Aut Ratio	0.60	6.09	0.46	0.10	0.07	–
Precipitation rate (mm day^{-1})	3.06	3.05	3.06	3.09	3.06	2.67
Cloud cover (%)	58.62	59.53	59.56	55.23	59.29	65.4
COT	15.2	15.2	15.3	9.5	15.9	13.1 ^a
CER (μm)	10.1	11.4	11.4	10.0	10.1	13.8 ^a
LWP (g m^{-2})	110	125	128	56	114	124 ^a
AOT	0.127	0.135	0.134	0.110	0.131	0.16 ^a
N_c burden (10^{10} m^{-2})	3.8	2.9	2.8	2.8	4.3	4.0 ^b
OLR (W m^{-2})	233.33	233.33	233.64	235.13	233.01	233.1
OSR (W m^{-2})	–230.76	–230.66	–233.40	–242.69	–229.06	–
OLR + OSR (W m^{-2})	2.57	2.67	0.24	–7.56	3.95	–
LWCRF (W m^{-2})	20.1	20.0	19.8	19.0	20.3	26.0
SWCRF (W m^{-2})	–56.9	–57.1	–57.4	–46.5	–58.5	–47.2
NetCRF (W m^{-2})	–36.8	–37.1	–37.6	–27.4	–38.1	–21.1

^a Averaged from 60°S to 60°N

^b AVHRR retrieval by Han et al. (1998) from 50°S to 50°N

3.3 Results

3.3.1 Global Statistics

The global mean results of climatological simulations from five schemes are summarized in Table 3.2. The satellite data products for comparison with the model outputs were obtained from: GPCP version 2.2 (Adler et al., 2003) for precipitation rate for the period 2001–2013; International Satellite Cloud Climatology Project (ISCCP) D2 dataset (Rossow and Schiffer, 1999) for cloud cover for the period 2005–2009; Aqua-MODIS level 3 monthly product (Hubanks et al., 2008) for cloud optical thickness (COT), cloud droplet effective radius at the cloud-top (CER), liquid water path (LWP), and aerosol optical thickness (AOT) for the period 2006–2010; NOAA interpolated outgoing longwave radiation (OLR) monthly product (Liebmann and Smith, 1996) for the period 2006–2010; CERES EBAF-TOA Edition 2.8 (Loeb et al., 2009) for shortwave/longwave/net cloud radiative forcing (SWCRF, LWCRF, NetCRF) at the top of the atmosphere

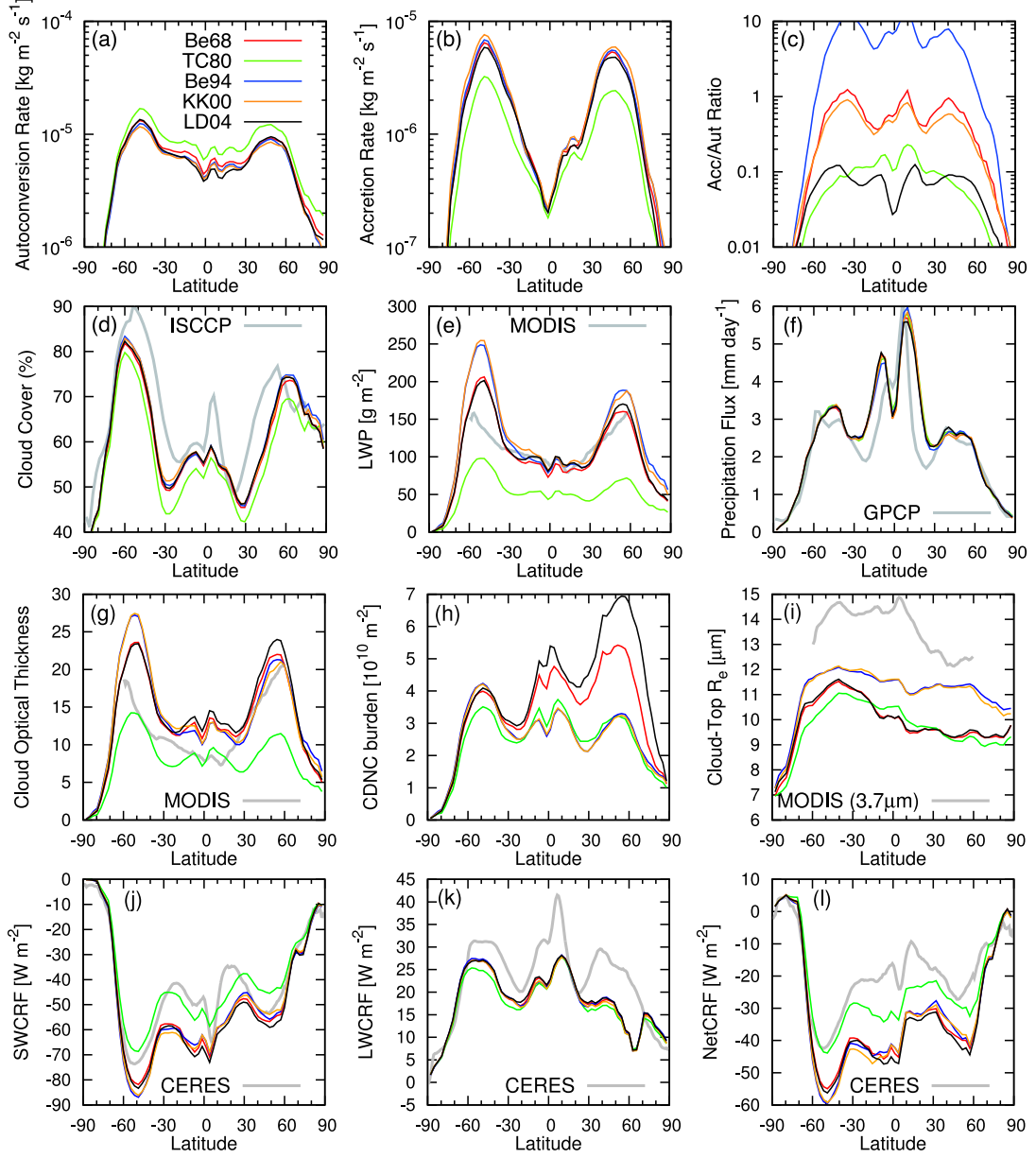


Fig. 3.2: Zonal mean (a) autoconversion rate, (b) accretion rate, (c) the ratio of accretion to autoconversion rate (Acc/Aut ratio), (d) total cloud cover, (e) liquid water path, (f) precipitation flux, (g) cloud optical thickness, (h) column integrated N_c , (i) cloud effective particle radius (R_e) at the cloud top, (j) shortwave cloud radiative forcing, (k) longwave cloud radiative forcing, and (l) net cloud radiative forcing, for five schemes (colored lines). Observational data from ISCCP D2 (for the period 2005–2009), Aqua-MODIS Level 3 (for the period 2006–2010), GPCP version 2.2 (for the period 2001–2013), and CERES EBAF–TOA Ed2.8 (for the period 2001–2013) are also plotted (gray lines) for reference (see text for details).

(TOA) for the period 2001–2013. Note that the MODIS-CER used here is obtained from 3.7 μm band (not from 2.1 μm band of the standard effective radius), because the 3.7 μm channel

represents upper cloud-top property (Platnick, 2000), which is better for the comparison with the model showing CER.

Cloud physical parameters of the total cloud cover, COT, CER, and LWP are similar between Be94 and KK00; however, the radiative parameters of the outgoing longwave/shortwave radiation (OLR, OSR) are slightly different. A similar tendency was found in the comparison between Be68 and LD04; although the COT, CER, and LWP are similar, radiative parameters have a high sensitivity, which results in differences of $> 1 \text{ W m}^{-2}$ for both the balance of OLR and OSR ($\text{OLR} + \text{OSR}$) and SWCRF. LWCRF is insensitive compared to SWCRF when changing the schemes. The result with the scheme of TC80 generally has a large gap compared to the results from the other schemes (e.g., cloud cover, COT, LWP, precipitation rate, OSR, and SWCRF). TC80 predicts a higher conversion rate and precipitation flux than any other scheme, resulting in the lowest LWP, cloud cover, and COT. As a result of the cloud–radiation interaction, the highest OLR and the smallest SWCRF are also observed in TC80.

Figure 3.2 shows the results of the zonal mean of the vertically integrated process rate (unit: $\text{kg m}^{-2} \text{ s}^{-1}$) and the ratio of accretion to autoconversion (Acc/Aut ratio) (Figs. 3.2a–3.2c), micro- and macrophysical parameters (Figs. 3.2d–3.2i), and the cloud radiative effects (Figs. 3.2j–3.2l) with satellite-retrieved data as references. Because the satellite data with regard to microphysical process rates (i.e., autoconversion and accretion rates) are limited for long-term and global-scale statistics, observational data are not presented in Figs. 3.2a–3.2c. Both autoconversion and accretion rates vary within 10^{-6} to $10^{-7} \text{ kg m}^{-2} \text{ s}^{-1}$, and their differences among the autoconversion schemes are obvious over the midlatitudes, especially in the accretion rate. The autoconversion rate is approximately half an order of magnitude or more higher than the accretion rate, and the Acc/Aut ratio ranges from 0.01 to 10.0 (Fig. 3.2c). It should be noted that the Fig. 3.2c has different meaning from just the ratio of Fig. 3.2b to Fig. 3.2a, because the Acc/Aut ratio is calculated in each time step (i.e., “instantaneous ratio”) and then the ratio averaged (see Gettelman et al., 2013, 2015). The Acc/Aut ratio is one of the key parameters for examining the microphysical conversion process (e.g., Jiang et al., 2010; Terai et al., 2012; Gettelman et al., 2013), and the large differences almost over the globe (i.e., with wide ranges from ~ 0.1 of Kessler-type schemes to 10.0 of model-based schemes) would have a high impact on rain formation. A detailed discussion about the process rates for horizontal/vertical distributions are provided later (cf. Section 3.3.2).

Precipitation flux is almost the same among the five schemes as shown in Table 3.2 and

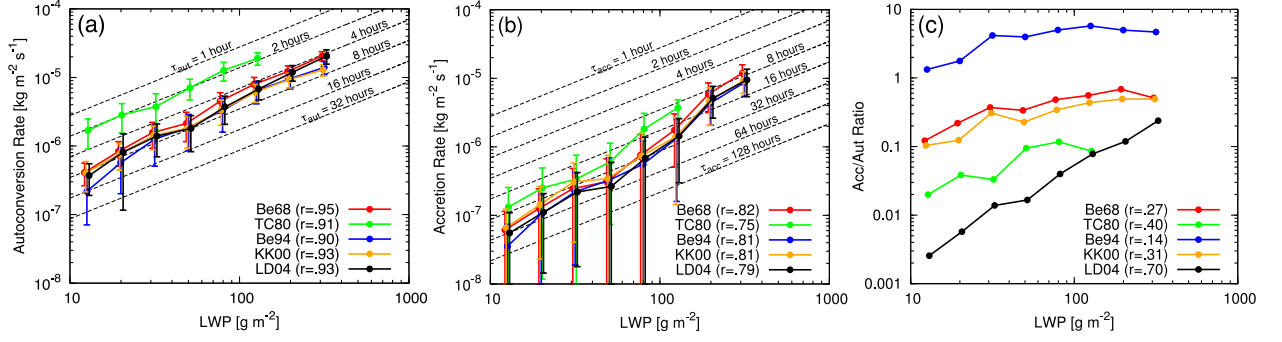


Fig. 3.3: The relationships between LWP and vertically integrated (a) autoconversion rate ($\text{kg m}^{-2} \text{s}^{-1}$), (b) accretion rate ($\text{kg m}^{-2} \text{s}^{-1}$), and (c) Acc/Aut ratio. The dashed lines in (a) and (b) represent the conversion timescales for each process, τ_{aut} and τ_{acc} , respectively. The correlation coefficient r between the variables of the abscissa and ordinate is presented in the legend.

Fig. 3.2f. This is because the simulations were run with the climatological sea surface temperature, which governs the evaporation rates from the sea surface. The equilibrium of the amount of evaporation and precipitation is not influenced in any of the simulations, and hence the differences in precipitation rate are insensitive. Although the difference in cloud cover is also relatively small, LWP and COT are very sensitive to the choice of schemes. More specifically, the scheme of TC80, which generates the highest amount of the rainwater, predicts the lowest cloud cover, LWP, and COT. These differences in cloud micro- and macrophysical properties also affect the AOT to some extent via the wet-deposition process. With respect to cloud microphysical parameters, both column integrated N_c and CER (Figs. 3.2h and 3.2i) are also highly sensitive to the scheme used. Column droplet number was estimated to be $\sim 4.0 \times 10^{-10} \text{ m}^{-2}$ from AVHRR satellite observation by Han et al. (1998) for an area between 50°S and 50°N . Although LD04 and Be68 schemes represent a little higher N_c burden than the other schemes, the model shows roughly similar results with Han et al. (1998). On the other hand, all schemes underestimate CER in a few μm , as compared with MODIS retrieval. Finally, as a result of the cloud-radiation interaction, the difference in NetCRF can be 10 W m^{-2} (from -27.4 to -38.1 W m^{-2}) depending on the autoconversion scheme.

3.3.2 Process Rate and Rain Formation

The parameterizations of process rates play important roles in the onset of precipitation and the global water cycle via the conversion processes, which generate the rainwater from cloud water. Figure 3.3 shows the variation in the process rates and the Acc/Aut ratio as a function of LWP.

This diagram has the advantage that it can display the autoconversion and accretion timescales (τ_{aut}, τ_{acc}) in the same dimension, as shown with dashed lines (Figs. 3.3a and 3.3b).

In the MIROC-SPRINTARS simulation, the autoconversion rate is predicted to be an order of magnitude higher than the accretion rate. The autoconversion rate increases by ~ 1 or 1.5 orders of magnitude as the LWP increases from ~ 10 to $\sim 300 \text{ g m}^{-2}$, with a more robust correlation than the accretion process, and the accretion rate increases by ~ 2 orders of magnitude when the LWP increases from ~ 10 to $\sim 300 \text{ g m}^{-2}$. Although the τ_{aut} is almost a constant value for any LWP value, the τ_{acc} becomes shorter with increasing LWP, especially at higher LWPs (i.e., $> 100 \text{ g m}^{-2}$). The behavior of these process rates results in a range of Acc/Aut ratio from 0.01 to 10.0, which monotonically increases with increasing LWP. This can be interpreted as an enhancement of the contribution of accretion compared to the autoconversion process, and its dependence on LWP (i.e., slope) is quite different with regard to the scheme type (i.e., model-base or Kessler-type schemes). More specifically, Kessler-type schemes (TC80 and LD04) underpredict the Acc/Aut ratio, and this can be because they ignore the lower autoconversion rate by their Heaviside step function. On the other hand, Be94 scheme shows higher Acc/Aut ratio because of its high nonlinear dependence (i.e., $\alpha = 4.7$ and $\beta = -3.3$). That is, Be94 predicts very small autoconversion rate at low L_c or at high N_c as shown in Fig. 3.1. This causes the overpredict of the Acc/Aut ratio. These differences in the Acc/Aut ratio among the schemes would cause large gaps in any cloud/rainwater diagnosis.

Figure 3.4 shows the global distribution of the vertically integrated autoconversion rate (unit: $\text{kg m}^{-2} \text{ s}^{-1}$) and Acc/Aut ratio. Although the pattern of the horizontal distribution is similar among the five schemes, both for the autoconversion rate and Acc/Aut ratio, the contrast is clearly different, especially over the North Pacific, South Pacific, Southern Indian Ocean, and North Atlantic regions where the LWP is relatively high. In these regions, both the simulated autoconversion rate and Acc/Aut ratio are high, and this tendency is consistent with some previous studies (Jiang et al., 2010; Gettelman et al., 2013, 2015), which demonstrated a robust relationship between LWP and Acc/Aut ratio. This relationship suggests that the autoconversion is dominant (onset of precipitation) in the case of lower LWP, which results in the lower Acc/Aut ratio. On the other hand, in the case of moderate or higher LWP, the relative contribution switches to the accretion regime and results in the high ratio.

Zonally averaged profiles of the layer mean autoconversion rate (unit: $\text{kg m}^{-3} \text{ s}^{-1}$) and Acc/Aut ratio are shown in Fig. 3.5. The highest autoconversion rate in any of the schemes is over

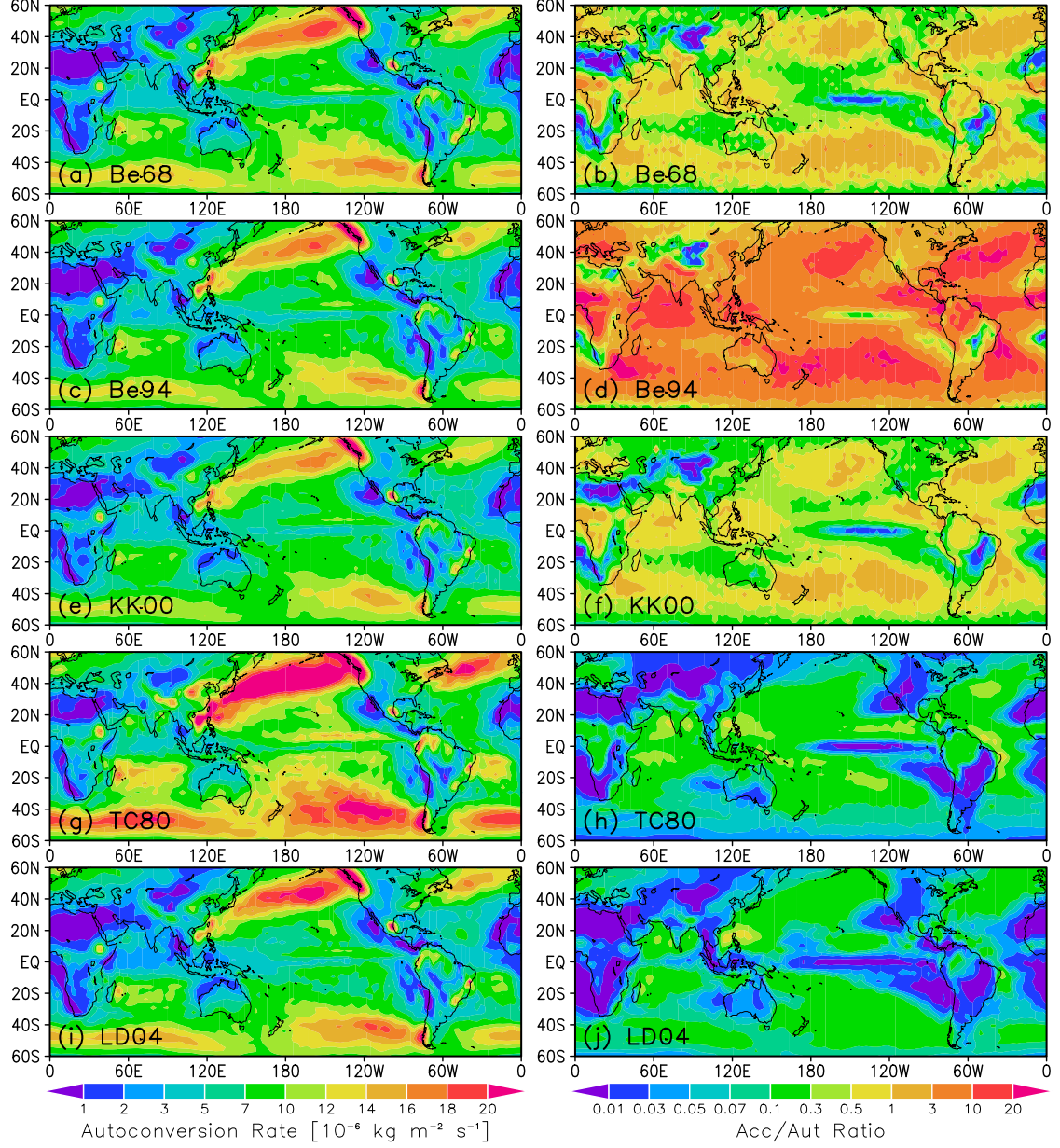


Fig. 3.4: The horizontal distribution of the vertically integrated (**left**) autoconversion rate ($\text{kg m}^{-2} \text{ s}^{-1}$), and (**right**) the Acc/Aut ratio simulated from the five schemes, (**a and b**) Be68, (**c and d**) Be94, (**e and f**) KK00, (**g and h**) TC80, and (**i and j**) LD04.

the region 40°S–50°S, around the level of 850 hPa, which corresponds to the South Pacific and Southern Indian Ocean as shown in Fig. 3.4. This tendency is also found over the northern mid-latitudes where the autoconversion rate is high (i.e., North Pacific). The rain formation process via autoconversion will strongly interact with ice and melting ice hydrometeors over these regions, hence future work will be required to investigate these processes including the ice physics. The

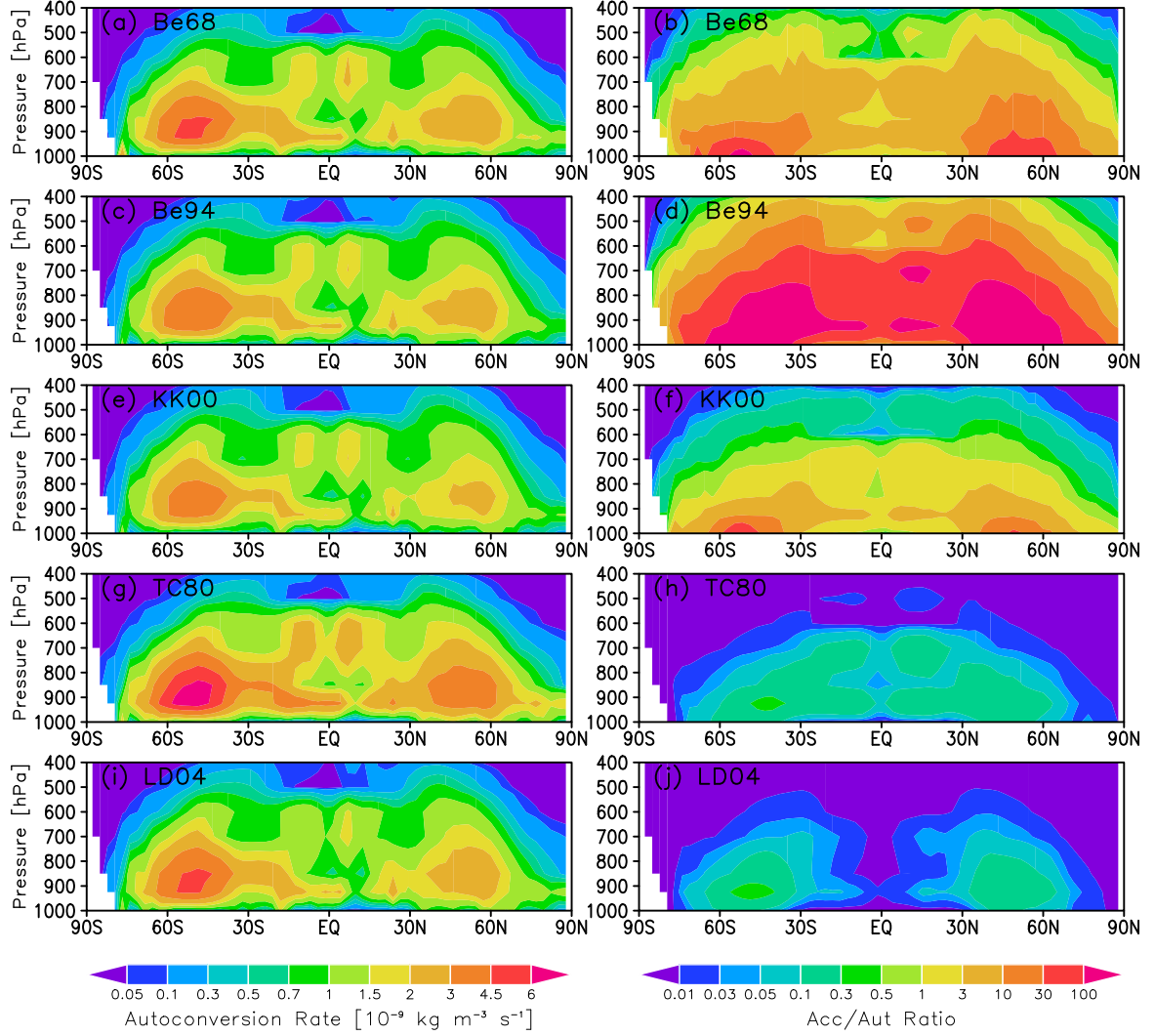


Fig. 3.5: Zonally averaged layer mean (**left**) autoconversion rate ($\text{kg m}^{-3} \text{ s}^{-1}$), and (**right**) the Acc/Aut ratio simulated from the five schemes, (**a and b**) Be68, (**c and d**) Be94, (**e and f**) KK00, (**g and h**) TC80, and (**i and j**) LD04.

vertical profile of the Acc/Aut ratio obviously differs depending on the autoconversion scheme, especially over the low-level midlatitudes, where most of the LSC is derived. More specifically, regions with a high Acc/Aut ratio (> 10.0) can be seen in the Be68, Be94, and KK00 schemes over the midlatitudes (i.e., more accretion); however, a lower Acc/Aut ratio (< 0.5) is dominant (i.e., more autoconversion) in the TC80 and LD04 schemes (with the Heaviside step function). Although the LD04 scheme predicts the lower Acc/Aut value than any other scheme, this behavior will strongly depend on the decision of the Heaviside step function. These differences regarding the vertical structure of the Acc/Aut ratio would result in large differences in rain formation

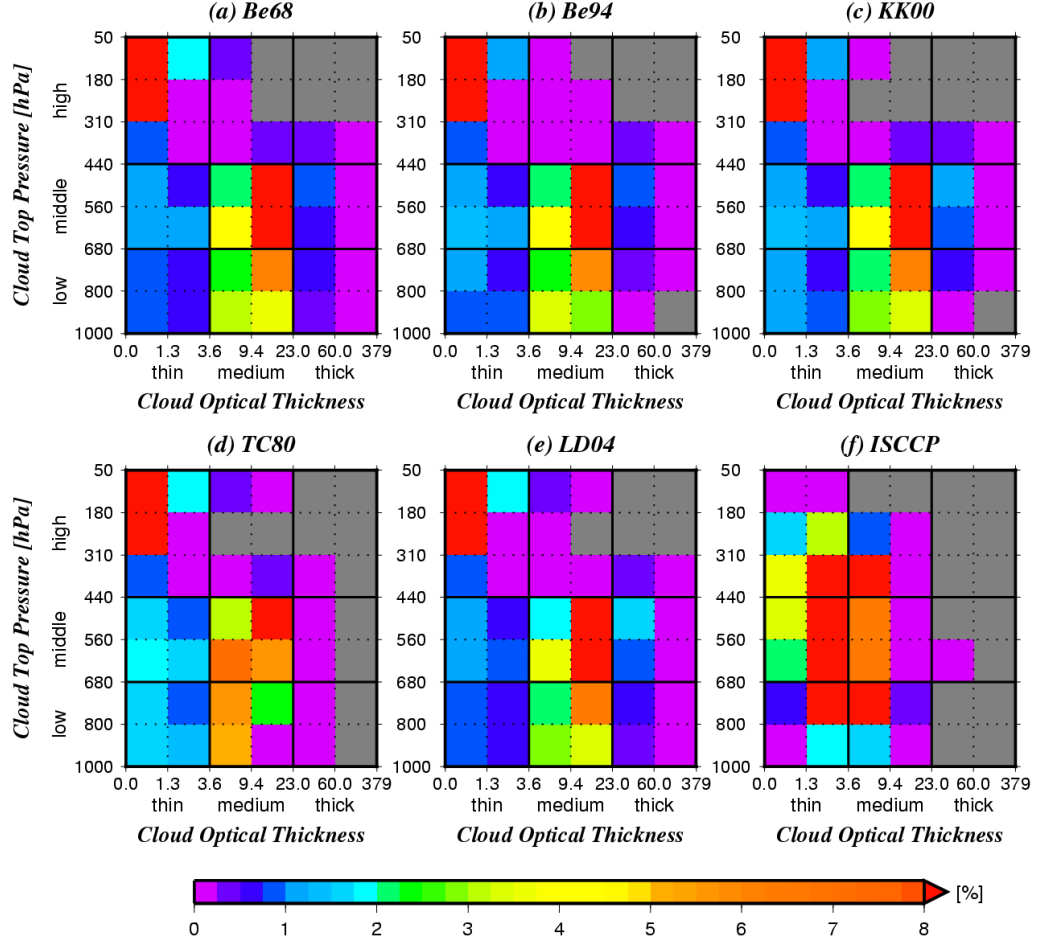


Fig. 3.6: Climatological cloud optical thickness (COT)–cloud top pressure (CTP) histogram proposed by Rossow and Stiffer (1999) for cloud-type definition. **(a–e)** MIROC-SPRINTARS simulations from different schemes, and **(f)** ISCCP D2 dataset. Over the tropics, the 25°N–25°S region was analyzed.

processes. This result is consistent with the previous studies which argued the importance of the choice of the autoconversion threshold (i.e., Heaviside step function for the onset of precipitation) (e.g., Pawlowska and Brenguier, 2003; Guo et al., 2008; Golaz et al., 2011, 2013).

Further discussion of the global radiation budget is provided in the next subsection in association with the cloud macrophysical vertical structure, which stems from the above differences in microphysical properties.

3.3.3 Cloud–Radiation Interaction

To determine the model uncertainty in cloud macrophysics, the ISCCP D2 dataset ($2.5^\circ \times 2.5^\circ$ gridded monthly mean product) was used for the period of 2005–2009 (i.e., 5 years). According to Rossow and Schiffer (1999), the COT range from 0 to 379 was divided into six bins with

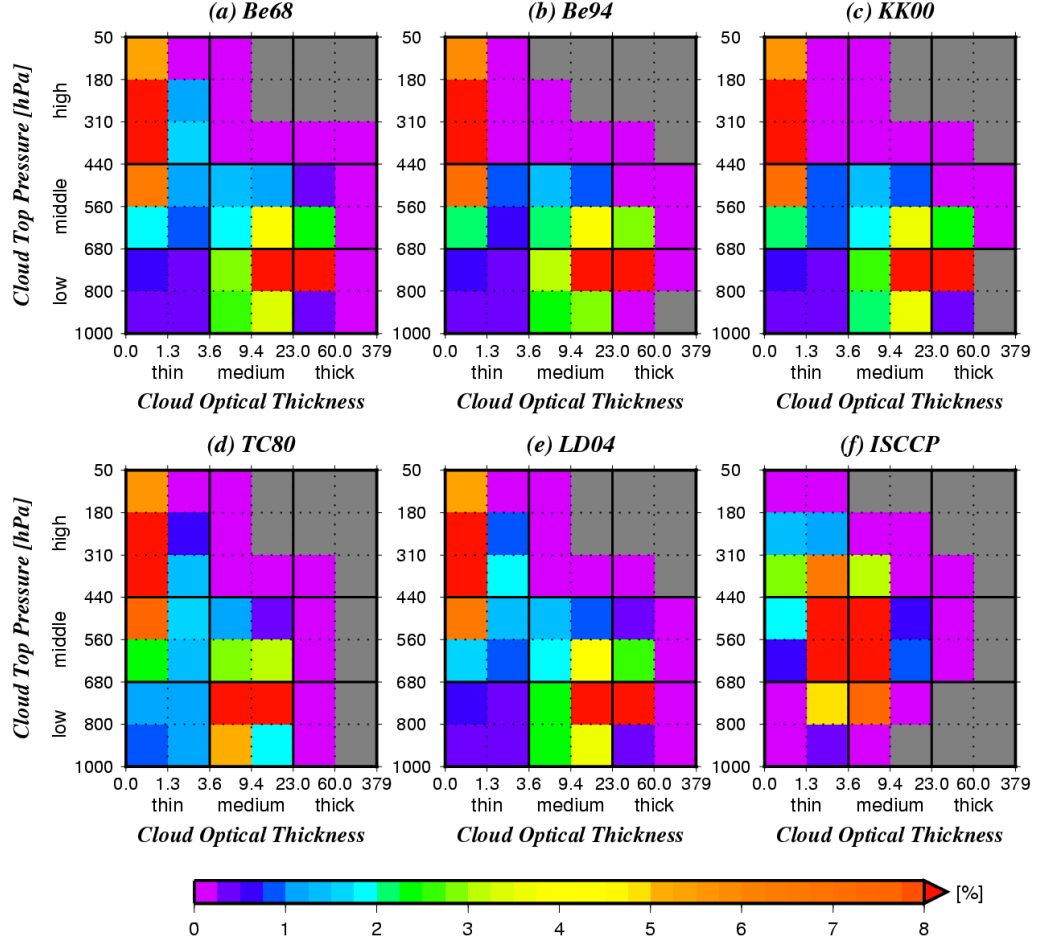


Fig. 3.7: Same as Fig. 3.6, but over the midlatitudes 25–50°N, 25–50°S.

thresholds of 1.3, 3.6, 9.4, 23, and 60, and the cloud top pressure (CTP) range from 1000 to 50 hPa was divided into seven bins with thresholds of 800, 680, 560, 440, 310, and 180.

The COT–CTP joint histograms are shown in Figs. 3.6 and 3.7 for the tropics and midlatitudes, respectively. Note that the histograms are obtained not from the ISCCP simulator but from the results of the model’s output. Many samples are concentrated at mid-levels, within the optically thin/medium categories in the observational data over the tropics region (Fig. 3.6f). The model generally simulates mid- and low-level clouds that are optically thicker than observations. A model bias also appears, with the high thin category (i.e., cirrus) having a too frequent prediction. These biases also appeared over the midlatitudes (Fig. 3.7). These discrepancies between observations and the model simulation were also reported in Kodama et al. (2012), who analyzed NICAM results with satellite simulators. A similar issue was also reported by Nam et al. (2012), who conducted an intercomparison of low clouds over the tropics among the fifth phase of the Coupled-Model Intercomparison Project (CMIP5) climate models with satellite simulators and

several satellite datasets. They suggested a tendency to underestimate the low cloud cover and overestimate its radiative effect in many of the GCMs in CMIP5. This so-called “too few, too bright low cloud problem (Nam et al., 2012)” is seen in MIROC-SPRINTARS as well (see also Figs. 3.2d and 3.2j).

Typical model errors of the cloud vertical structure, as shown in Figs. 3.6 and 3.7 and in the previous studies mentioned above, need to be improved for an accurate representation for cloud–precipitation–radiation interaction. In addition, COT and CTP are insensitive to the differences in autoconversion schemes, which suggests that the models need a more fundamental improvement than just in their microphysical parameterizations. To this end, new parameterizations for the convective cloud fraction (Gianotti and Eltahir, 2014a) and an autoconversion scheme for convective rainfall (Gianotti and Eltahir, 2014b) were recently proposed, and can explicitly consider the subgrid variability. Alternatively, the introduction of a prognostic rain scheme (e.g., Posselt and Lohmann, 2008, 2009; Wang et al., 2012; Gettelman and Morrison, 2015; Gettelman et al., 2015) will also lead to a more accurate reproduction of the global water cycle and radiation budget. Johnson et al. (2015) evaluated the uncertainty in convective cloud microphysics, and concluded that the collection efficiency of cloud drops by graupel particles was one of the largest factors in model uncertainty. Although the process of solid precipitation is beyond the scope of this study, which examines only the warm phase rain formation process, development of microphysical parameterization including not only liquid cloud but also ice/snow physics (Gettelman et al., 2010; Gettelman and Morrison, 2015) will be vitally important for future model improvement.

Figure 3.8 shows the relation between LWP and NetCRF from satellite observations (combined with MODIS-LWP and CERES-NetCRF for the period 2006–2010, i.e., 5 years) and model simulations for five autoconversion schemes. The NetCRF from both the observation and model results decrease monotonically with increasing LWP, and the correlation between LWP and NetCRF (r with the legend in the right-top of Fig. 3.8) is much stronger in model simulations ($r = \sim -.70$ or more robust) than the observation ($r = -.35$). The difference in NetCRF between the schemes is generally within a few W m^{-2} , except for TC80, which attains $\sim 5 \text{ W m}^{-2}$ or larger difference. However, the discrepancy between the observed and simulated NetCRF is larger (~ 20 or more W m^{-2} when LWP exceeds 100 g m^{-2}) than the above differences between the schemes, which suggests that a dramatic improvement of radiative properties cannot be anticipated by changing just the microphysical (i.e., autoconversion) scheme. The overestimation of NetCRF would be

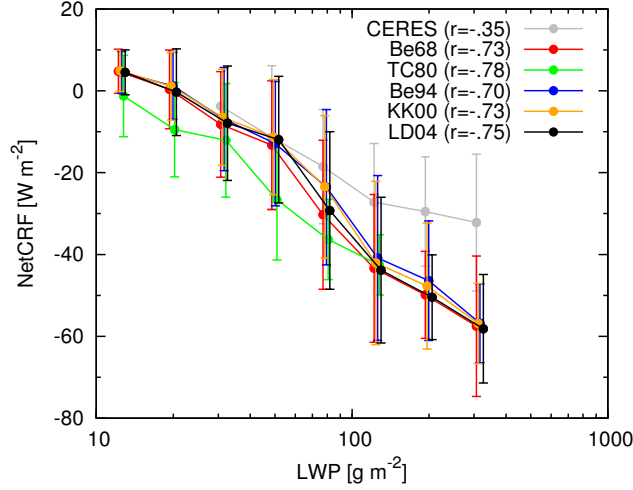


Fig. 3.8: The relationships between LWP and the net cloud radiative forcing (NetCRF; i.e., SWCRF + LWCRF) from CERES observation (gray) and MIROC-SPRINTARS simulation (colored). The correlation coefficient (r) between LWP and NetCRF is presented with a legend.

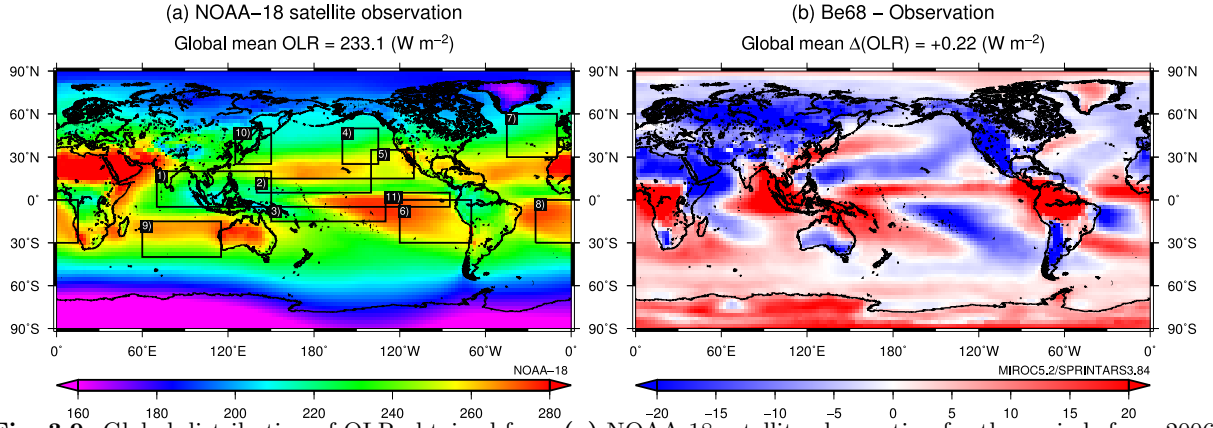


Fig. 3.9: Global distribution of OLR obtained from (a) NOAA-18 satellite observation for the periods from 2006 to 2010 (i.e., five years), and (b) the anomaly of Berry (1968) from NOAA. The spatial resolution in NOAA-18 is 2.5° , and is down-scaled to 2.8° when subtracted from the model simulations. The grid-boxes and white numbers correspond to Table 3.3 for regional analysis (cf. Section 3.3.4).

caused by overestimating COT in low clouds. The direct cause of the overestimating COT is probably because of the underestimating CER in a few μm as mentioned earlier (Table 3.2 and Fig. 3.2i), and hence SWCRF would be overpredicted. Furthermore, N_c is also important with respect to the microphysical aspect, for example, the LD04 scheme predicts the highest N_c burden, which results in the largest SWCRF and COT. Although the Be68 scheme also shows higher N_c and smaller CER, SWCRF seems to be offset by the underestimate of cloud cover to some extent.

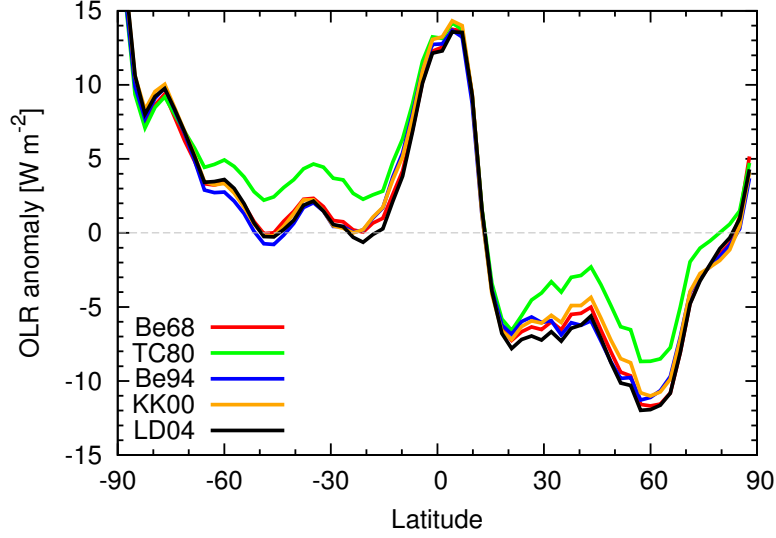


Fig. 3.10: Zonal mean OLR anomaly (W m^{-2}) from NOAA-18 satellite observations.

Figure 3.9a shows the global OLR distribution obtained from NOAA-18 satellite observations ($2.5^\circ \times 2.5^\circ$ in latitude and longitude) for 5 years (2006–2010), and Fig. 3.9b shows the anomaly of the default simulation with Be68 from the NOAA-18 observation. The model spatial resolution is $2.8^\circ \times 2.8^\circ$ in latitude and longitude as described earlier, thus the satellite data were down-scaled when subtracted from the model results. The simulated global mean OLR differs from the observation by $+0.22 \text{ W m}^{-2}$. The model performs well quantitatively, but in terms of its distribution, the simulated result is not entirely satisfactory, especially the positive biases over the tropical convective region, and the negative biases over land. The zonal mean OLR anomaly from satellite observation is shown in Fig. 3.10. It is noteworthy that the difference from each scheme is smaller over low latitudes and conversely larger over midlatitudes, which is mostly caused by LSC in the warm rain microphysics, based on the autoconversion process. The difference in the radiative properties depending on the autoconversion schemes in Fig. 3.8 over the midlatitudes suggests that such biases in the model originate from the cloud micro- and macrophysical vertical structure as shown in Figs. 3.5 and 3.7.

3.3.4 Regional Analysis

The dependence of cloud characteristics on autoconversion schemes at the global or tropics/mid-latitude scale was discussed above. To understand the performance of the autoconversion parameterization in detail, a regional analysis was conducted. The main purpose of this was to evaluate

Table 3.3: Definition of the 11 regions and references for regional analysis, corresponding to the boxes shown in Fig. 3.9.

	Region	Latitude, Longitude	Reference
1)	Tropical Warm Pool	5°S–20°N, 70°E–150°E	Franklin et al. (2013a,b)
2)	ITCZ	5°N–15°N, 140°E–140°W	Kubar et al. (2009)
3)	SPCZ	15°S–5°S, 150°E–130°W	Kubar et al. (2009)
4)	NE Pacific	25°N–50°N, 160°W–135°W	Michibata et al. (2014)
5)	California	15°N–35°N, 140°W–110°W	Franklin et al. (2013a,b)
6)	Peruvian	30°S–0°S, 120°W–70°W	Leon et al. (2008)
7)	North Atlantic	30°N–60°N, 45°W–10°W	Leon et al. (2008)
8)	Namibian	30°S–0°S, 25°W–15°E	Leon et al. (2008)
9)	Australian	40°S–15°S, 60°E–115°E	Leon et al. (2008)
10)	Japan	25°N–50°N, 125°E–150°E	Michibata et al. (2014)
11)	Eqt. Cold Tongue	5°S–5°N, 130°W–85°W	Kubar et al. (2009)

how the reproducibility of cloud and radiative properties is influenced by the difference in the autoconversion schemes associated with atmospheric conditions (e.g., static stability, aerosol concentration). To this end, 11 regions were defined with reference to several previous studies as shown in Table 3.3. There were three tropical regions: **1)** Tropical Warm Pool, **2)** Inter-Tropical Convergence Zone (ITCZ), and **3)** South Pacific Convergence Zone (SPCZ). These regions significantly influence the global water cycle through deep convection and experience large amounts of precipitation. There were six west coastal stratocumulus regions: **4)** NE Pacific, **5)** California, **6)** Peruvian, **7)** North Atlantic, **8)** Namibian, and **9)** Australian, where optically bright low clouds spread. The region of **10)** Japan is an outflow region of anthropogenic aerosols from industrial East Asia, and region **11)** Equatorial Cold Tongue is a comparably stable environment due to an inflow of cold current from off the coast of Peru.

Figure 3.11 shows a scatterplot as a function of LWP and autoconversion rate, and was used to examine how the liquid cloud formation varies due to the different autoconversion schemes. Although the global statistics show that the TC80 has the highest autoconversion rate, followed by Be68 and LD04 (see Fig. 3.3a and Table 3.2), LD04 predicts the lowest autoconversion rate over the tropics (Figs. 3.11a–3.11c). This complex behavior would be involved in two reasons: (1) the differences in N_c among regions bring about changes in the autoconversion rate and its timescale; (2) the differences in non-linear dependence of L_c and N_c (i.e., α and β in the equation (3.1)) on the autoconversion rate among the schemes contribute to determination of the slope. Basically, each scheme has similar relationship between LWP and the autoconversion rate (i.e., slope) in different regions. In general, TC80 has the fastest autoconversion timescale within the

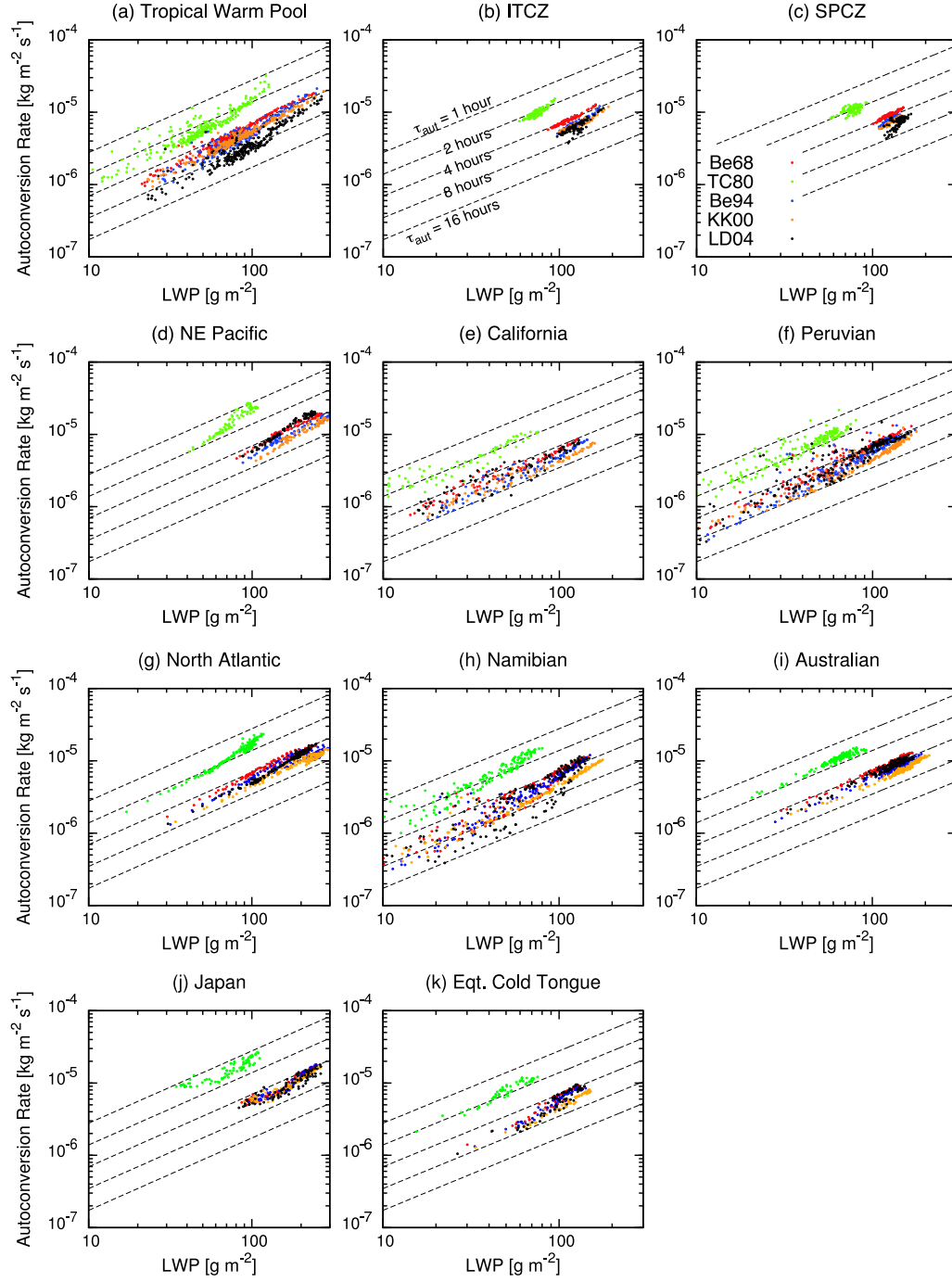


Fig. 3.11: Scatterplot of LWP vs. autoconversion rate for each region from five different autoconversion schemes. The dashed lines represent the autoconversion timescale (τ_{aut}) of 1 to 16 h.

dashed line of 1–2 hours, which is about 2 or more times faster than the autoconversion timescale of the other schemes in any region, such as the high autoconversion rate and fast conversion time resulting in the generation of large amounts of precipitation, as described earlier.

The NE Pacific and North Atlantic (Figs. 3.11d and 3.11g) are regions with a high occurrence of drizzle, and Japan in the coastal region of East Asia (Fig. 3.11j) has a relatively high LWP in simulations, which is consistent with previous studies that used satellite retrievals (Leon et al., 2008; Michibata et al., 2014). There are also noteworthy differences between the NE Pacific (Fig. 3.11d) and California (Fig. 3.11e). Although these two regions are located a very short distance from each other, a lower conversion rate and LWP are simulated over California. California is a typical stratocumulus region, characterized by a high low cloud fraction and lower radar reflectivity (Leon et al., 2008; Franklin et al., 2013a). The model predicts a slower autoconversion timescale in this region, which does not conflict with the observations. Such ‘calm’ characteristics, with a low LWP and autoconversion rate, are also seen in other stratocumulus regions, i.e., the Peruvian and Namibian regions (Figs. 3.11f and 3.11h), where the autoconversion timescale is very long (most samples are concentrated in ~ 4 –8 hours). The difference in the autoconversion timescale due to changing autoconversion scheme is quite large, which will affect cloud lifetime and water cycle.

In terms of the cloud–radiation interaction, the relationship between LWP and NetCRF was examined again, as shown in Fig. 3.12. NetCRF monotonically decreases with increasing LWP, as expected. However, the quantitative radiative effect is overestimated compared to the CERES observation in most regions, as also shown in Fig. 3.8 for the global statistics. This feature was also reported in most of the CMIP5 models (e.g., Nam et al., 2012; Wang and Su, 2013; Lauer and Hamilton, 2013; Calisto et al., 2014).

In general, the model simulates a wide range of LWPs, whereas satellite results are concentrated in a narrow range. Over the California, Peruvian, North Atlantic, and Japan regions (Figs. 3.12e, 3.12f, 3.12g, and 3.12j) the relation between LWP and NetCRF is relatively well predicted, but there are errors in NetCRF to some extent (~ 5 or more W m^{-2}). In the tropics, the SPCZ region (Fig. 3.12c) has a particularly large uncertainty. We should note that the comparison between model and observation over the tropical regions is difficult because MODIS cannot accurately retrieve LWP for high cloud and deep convective cloud with ice phase. The Namibian and Equatorial Cold Tongue regions (Figs. 3.12h and 3.12k) also have some uncertainty that could not be compensated for by just changing the autoconversion scheme.

Overall, the importance of choosing the appropriate autoconversion scheme for reliable climate prediction is highlighted by the regional analysis described above, and these results also suggest that there are still more fundamental errors without microphysical process in the models.

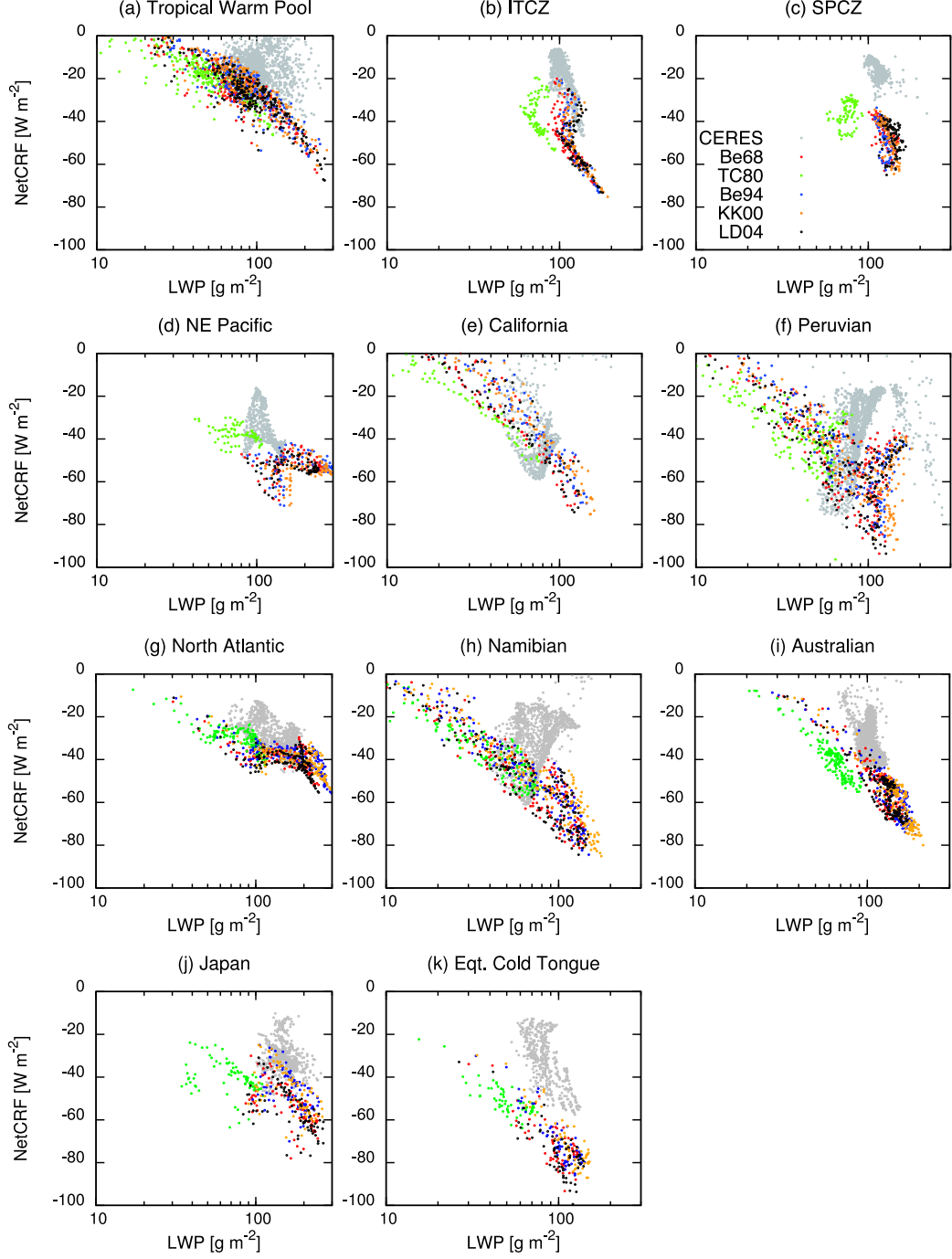


Fig. 3.12: Scatterplot of LWP vs. NetCRF for each region from five different autoconversion schemes. The observation data combined with MODIS-LWP and CERES-NetCRF (for the period 2006–2010) is also plotted in gray as a reference.

3.4 Summary Perspectives

We conducted sensitivity experiments with five commonly used autoconversion schemes using the aerosol–climate model MIROC-SPRINTARS to evaluate the performance of the schemes. The

potential uncertainties in the cloud physical process and radiative properties were also examined using several satellite datasets (MODIS, ISCCP, GPCP, CERES, and NOAA-18).

In terms of rain formation processes, the TC80 autoconversion scheme had a higher autoconversion rate than the other four schemes examined in this study, which resulted in the generation of a large amount of precipitation. The Acc/Aut ratio, which is a key parameter for examining the balance of microphysical conversion process from cloud water to rainwater (e.g., Jiang et al., 2010; Terai et al., 2012; Gettelman et al., 2013), was clearly different globally when the autoconversion scheme was switched. The Acc/Aut ratio monotonically increased with increasing LWP in any schemes, but its dependence on LWP was quite different. It is noteworthy that Kessler-type schemes underestimated the Acc/Aut ratio, and this would be primarily related to the autoconversion threshold (e.g., Guo et al., 2008; Golaz et al., 2011). The vertical profile of the Acc/Aut ratio was also highly sensitive to a change in the autoconversion scheme, especially over the low-level midlatitudes where the LSC makes its main contribution. Therefore, the large differences in the Acc/Aut ratio over the midlatitudes had a large impact on rain formation.

In contrast, Gettelman et al. (2013) estimated that the Acc/Aut ratio ranges from 1 to ~ 100 from VOCALS aircraft measurements (Wood et al., 2011; Terai and Wood, 2013). MIROC-SPRINTARS simulated the Acc/Aut ratio from 0.01 to 10.0 as shown in Fig. 3.3, which means that the relative role of the accretion process is one or more orders of magnitude underestimated compared to the observation. Posselt and Lohmann (2008) suggested that the diagnostic rain schemes tend to depend excessively on the autoconversion process, hence not only MIROC-SPRINTARS, but also practically all GCMs that apply a diagnostic treatment of rain will have the above issue. To overcome the issue concerning the underestimation of Acc/Aut ratio, a prognostic rain scheme would be an improvement in GCMs. This method can keep the rainwater within the atmosphere for two or more time steps, thus the contribution of the accretion process is enhanced (Posselt and Lohmann, 2008; Gettelman and Morrison, 2015). In fact, recent studies with the prognostic rain scheme successfully represented a more realistic balance of precipitation processes (Gettelman et al., 2015), and reduced the occurrence of spurious light rain (Walters et al., 2014).

The simulated cloud macrophysical structure was compared to ISCCP satellite retrieved data using COT-CTP joint histograms. The model generally predicted low clouds that were optically too bright, and frequently overestimated high thin clouds. This poor representation of cloud vertical structure affects the radiative properties. The difference between the observed and sim-

ulated NetCRF was quite large (~ 20 or more W m^{-2}) when LWP exceeds 100 g m^{-2} . These important issues concerning the cloud vertical structure and overestimating the cloud radiative effect, which are known to be typical of most GCMs, were found in MIROC-SPRINTARS as well as by comparison with satellite observations, yet they could not be overcome by changing just the autoconversion scheme. These results strongly suggest that models need fundamental improvements, not only in their microphysical parameterization (i.e., autoconversion), but also in their dynamical/macrophysical parameterizations, for the better prediction of the cloud–precipitation–radiation interaction. For example, introducing a triclass parameterization including drizzle (i.e., cloud water, drizzle, and rainwater) would represent a more realistic precipitation process (Sant et al., 2013).

Nevertheless, a detailed regional analysis confirmed that the difference in the autoconversion schemes definitely affected the conversion timescale from cloud water to rainwater, and the radiative properties, hence the appropriate choice of microphysical parameterization is very important for the performance of climate models as a whole. In fact, recent studies assessed the relationship between the nonlinear dependence of autoconversion schemes and indirect aerosol effect from single-column model study (Suzuki et al., 2013a) and multimodel framework (Wang et al., 2012; Suzuki et al., 2015, in press). Furthermore, Ma et al. (2015) investigated the resolution dependence of the aerosol-cloud-precipitation interaction, and demonstrated that the higher-resolution model can reproduce more realistic microphysical susceptibility (lower indirect aerosol effect). The simulation with higher resolution and shorter time step will gain a better understanding of the microphysical conversion processes. The process of autoconversion plays a particularly important role in the determination of subgrid-scale circumstances due to its high nonlinearity. For example, new parameterizations for the convective cloud fraction (Gianotti and Eltahir, 2014a) and autoconversion schemes for convective rainfall (Gianotti and Eltahir, 2014b) that can explicitly account for subgrid variability were proposed. In the real atmosphere, the atmospheric dynamical conditions (e.g., static stability, vertical velocity, and boundary layer depth) or the aerosol concentration will fluctuate within a subgrid-scale, hence leading to uncertainty due to the difficulty in representing their behavior by the current parameterization, with the coarser resolution of GCMs.

The critical issue regarding the treatment of subgrid-scale variability is not limited to MIROC-SPRINTARS, but also applies to almost all GCMs. In fact, recent studies have stressed the importance of the treatment of subgrid-scale variability in the rain formation process (Weber and

Quaas, 2012; Turner et al., 2012; Tonttila et al., 2015). A future examination of the dependency of the rain formation process on the model resolution (e.g., Li et al., 2015) will provide important insights. Furthermore, the effect of turbulence on cloud processes and precipitation should be considered (Franklin, 2008; Wyszogrodzki et al., 2013), and warm rain schemes including such interactions have been developed (Franklin, 2014; Grabowski et al., 2015).

In this study, the dependency of the performance of autoconversion schemes on spatial/temporal resolution was not considered, therefore future work will investigate this issue. As discussed above, the results and recommendations from this evaluation of autoconversion schemes, and the potential uncertainty in cloud physical process through the sensitivity tests using only one specific model, will contribute to more accurate model development.

4 Unveiling the Source of Discrepancies between GCMs and A-Train in Aerosol-Cloud-Precipitation Interactions

Abstract

Aerosol–cloud interactions are one of the most uncertain processes in climate models due to their nonlinear complexity. A key complexity arises from the possibility that clouds can respond to perturbed aerosols in two opposite ways, as characterized by the traditional “cloud lifetime” hypothesis and more recent “buffered system” hypothesis. Their importance in climate simulations remains poorly understood. Here we investigate the response of the liquid water path (LWP) to aerosol perturbations for warm clouds from the perspective of process-oriented model evaluations using a general circulation model (GCM) and A-Train satellite products. A systematic difference is found in the LWP response between the GCM results and observations. The model results indicate a near-global uniform increase of LWP with increasing aerosol loading, while the sign of the response of the LWP from the A-Train varies from one region to another. The satellite-observed response of the LWP is closely related to meteorological and/or macrophysical factors, in addition to the microphysics. The model does not reproduce this variability of cloud susceptibility (i.e., sensitivity of LWP to perturbed aerosols) because the parameterization of the autoconversion process assumes only suppression of rain formation in response to increased cloud droplet number, and does not consider macrophysical aspects that serve as a mechanism for the negative responses of the LWP via enhancements of evaporation and precipitation. Model biases are also found in the precipitation microphysics, which suggests that the model generates rainwater readily even when little cloud water is present. This essentially causes projections of unrealistically frequent and light rain, with high cloud susceptibilities to aerosol perturbations.

Keywords: aerosol–cloud interactions, GCM biases, buffered system, climate

Key Points:

- The main source of model uncertainties is the response of the liquid water path (LWP) to aerosol perturbations.

- Observations clearly show both an increase and decrease in the response of LWP, depending on the exact macrophysical regime.
 - Our process-oriented comparison between GCM results and observations will enable the development of more accurate models.
-

4.1 Introduction

Aerosol particles play an important indirect role in the climate system by modifying cloud micro- and macrophysical properties, which is referred to as aerosol–cloud interactions (Twomey, 1977; Albrecht, 1989). An increase in aerosols supplies more numerous cloud condensation nuclei, resulting in numerous and smaller cloud droplets leading to brighter clouds, which is known as the “albedo effect” (Twomey, 1977). Smaller cloud droplets suppress the onset of precipitation in warm clouds due to the less efficient collision–coalescence process, resulting in a longer cloud lifetime, which is known as the “lifetime effect” (Albrecht, 1989). There has been much discussion about the climatic impacts of aerosol-induced modulation of water clouds, which are particularly sensitive to aerosol perturbations (e.g., Pincus and Baker, 1994; Bréon et al., 2002; Penner et al., 2006; Lebsock et al., 2008; Quaas et al., 2009; Terai et al., 2012, 2015). However, quantitative estimates of radiative forcing with regard to aerosol–cloud–precipitation–climate interactions remain uncertain, as reported in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (2013).

One of the most important factors that quantify the magnitude of aerosol–cloud interactions is the response of the cloud liquid water path (LWP) to aerosol perturbations. This factor also characterizes aerosol impacts on the global hydrological cycle through its representation of the aerosol effect on precipitation efficiency. This effect is represented in general circulation models (GCMs) as aerosol-induced changes in rainwater production from cloud water, which are parameterized with a bulk microphysics as the so-called autoconversion process. The water conversion rate by this process (P_{aut}) is generally given as a function of the liquid water content (L_c) and cloud droplet number concentration (N_c) as

$$P_{\text{aut}} \sim L_c^\alpha \times N_c^{-\beta}, \quad (4.1)$$

where α and β are prescribed constants (e.g., Berry, 1968; Beheng, 1994; Khairoutdinov and Kogan, 2000). The N_c is then somehow related to the aerosol number concentration (N_a). In

GCMs, Eq. (4.1) provides the only pathway through which aerosols modulate precipitation formation and, thus, the cloud lifetime. Note that GCMs also partly include the opposing processes (decreasing LWP due to enhancement in evaporation) via the so-called direct and semi-direct effect (e.g., Ackerman et al., 2000; Hansen et al., 1997). Given that rainwater production is always suppressed with increasing N_c according to Eq. (4.1), GCMs tend to increase the LWP uniformly with increasing N_c for stratiform clouds.

On the other hand, some observational studies have shown two pathways of LWP responses to perturbed aerosols, i.e., both increasing and decreasing tendencies of LWP with increasing aerosols (Matsui et al., 2006; Lebsock et al., 2008; Chen et al., 2014); the mechanisms for these opposing responses cannot be understood by a simple microphysical argument alone, but are likely to relate to macrophysical and meteorological factors as well (e.g., Ackerman et al., 2004; Matsui et al., 2006; Suzuki et al., 2008; Lebsock et al., 2008; Small et al., 2009; Gryspeerdt and Stier, 2012; Chen et al., 2014). Wood (2007) found that there are processes that modify the cloud geometric thickness to aerosol perturbations in such a way that cancels the aerosol indirect effect at sufficiently long timescales. Such a compensation mechanism is currently considered one of the “buffering effects” (Stevens and Feingold, 2009), which generate the opposite result to the original hypotheses of cloud albedo and lifetime effects, for the cloud system as a whole (Rosenfeld et al., 2014; Lebo and Feingold, 2014; Seifert et al., 2015). Despite its critical importance to accurate climate simulations, the operation of this mechanism at the global scale remains poorly understood.

To determine the mechanisms involved in the competition between the “lifetime effect” and “buffering effect”, the complexity in aerosol effects on clouds needs to be untangled at a fundamental process-level. For this purpose, GCMs should be evaluated extensively against observations in the context of their process representations, which are key to the aerosol–cloud–precipitation interaction.

In this study, we analyze results from both GCM and A-Train data, with a particular focus on their discrepancies in the key indices of aerosol–cloud interactions relating to fundamental processes. The factors examined are the susceptibilities of cloud optical thickness (τ_c), droplet effective radius (r_e), and LWP to N_c . To focus directly on the cloud physical parameters, we use N_c as an aerosol proxy rather than N_a (e.g. Koren and Feingold, 2011). A satellite-based study by Chen et al. (2014) reported that cloud susceptibilities show similar results whether aerosol index, aerosol optical depth (AOD), or N_c are applied as an aerosol proxy (see their supplementary

information). Given the fundamental relationship of $\tau_c \propto \text{LWP} / r_e$, the susceptibilities are related as follows (Ghan et al., 2016):

$$\frac{d \ln \tau_c}{d \ln N_c} = -\frac{d \ln r_e}{d \ln N_c} + \frac{d \ln \text{LWP}}{d \ln N_c}, \quad (4.2)$$

where the first and second terms on the right side of Eq. (4.2) represent the “albedo effect” and the “lifetime effect”, respectively. Equation (4.2) has the advantage that it can quantify the contributions from the two effects that determine the aerosol impact on cloud radiative properties. As discussed here and also in recent studies (Ghan et al., 2016; Feingold et al., 2016), the two terms in Eq. (4.2) are related to representations of different processes. This approach makes it easier to understand the mechanisms that determine the resultant magnitude of aerosol indirect forcing in the context of relevant processes (Seinfeld et al., 2016).

The aim of the study is to clarify the fundamental source of uncertainty in process representations of aerosol–cloud–precipitation interactions in GCMs for stratiform and shallow cumulative warm clouds (excluding deep convective thick clouds or ice clouds; see Sect. 2). Given that the aerosol–cloud interaction processes are also influenced by both macrophysics (e.g., environmental conditions, dynamical regime, cloud type) and microphysics, we also place an emphasis on the importance of macrophysics (e.g., Sorooshian et al., 2013; Gryspeerd et al., 2014; Zhang et al., 2016).

4.2 Data

4.2.1 MIROC–SPRINTARS

A global climate model, Model for Interdisciplinary Research On Climate (MIROC) version 5.2 (Watanabe et al., 2010), was used in this study. The interactions of the main tropospheric aerosols (i.e., black carbon, organic matter, soil dust, sea salt, sulfate, and the precursor gases of sulfate) with cloud–precipitation microphysics and radiation–climate effects are incorporated in the aerosol module, Spectral Radiation-Transport Model for Aerosol Species (SPRINTARS) (Takemura et al., 2000, 2002, 2005), which is coupled with MIROC (MIROC–SPRINTARS).

The cloud macro- and microphysics framework in MIROC–SPRINTARS is based on a prognostic large-scale condensation scheme, which explicitly considers subgrid-scale variability of clouds (Watanabe et al., 2009). This probability distribution function (PDF)-based prognostic cloud scheme couples with the ice microphysics scheme proposed by Wilson and Ballard (1999). MIROC–SPRINTARS treats both cloud droplets and ice crystals as a two-moment bulk micro-

physics scheme (Takemura et al., 2009). The nucleation of cloud droplets is parameterized by the scheme of Abdul-Razzak and Ghan (2000), and the process of cloud-to-rain water conversion is diagnosed based on the Berry (1968) autoconversion scheme. Rainwater is not a prognostic variable in the current version of MIROC-SPRINTARS.

We extracted warm-phase low clouds (> 273.15 K in whole cloud layers) from every 6 h instantaneous output for 5 full years; as a result, 1 595 753 warm cloud samples were obtained. The horizontal and vertical resolutions were T42 (approximately $2.8^\circ \times 2.8^\circ$ in latitude and longitude) and 20 layers, respectively. A more detailed description of the model and its settings are documented in Michibata and Takemura (2015).

4.2.2 CloudSat and MODIS

We used the synergistic satellite data sets of the CloudSat and MODIS, which are both part of the A-Train constellation (Stephens et al., 2002, 2008). The data products, 2B-TAU (Polonsky, 2008), 2B-GEOPROF (Marchand et al., 2008), and ECMWF-AUX (Partain, 2007) were used for the period June 2006 to April 2011, i.e., a total of 5 full years. This facilitated the construction of stable statistics with a horizontal resolution (2.5° grid boxes in this study) close to the GCM output. We defined the cloud layer as where the cloud mask value is greater than 30 from the 2B-GEOPROF product, which means a good, or strong, echo with high-confidence detection (Marchand et al., 2008). The analysis was restricted to single-layer water clouds; in total, 7 872 426 cloud samples were obtained.

The LWP was derived from the MODIS-retrieved optical thickness and effective radius using the following equation for an adiabatically stratified cloud (Szczodrak et al., 2001):

$$\text{LWP} = \frac{5}{9} \tau_c r_e. \quad (4.3)$$

N_c was also calculated based on an adiabatic assumption (Wood, 2006) as

$$N_c = \sqrt{2} B^3 \Gamma_{\text{eff}}^{1/2} \frac{\text{LWP}^{1/2}}{r_e^3}, \quad (4.4)$$

where $B = (3/4\pi\rho_w)^{1/3} = 0.0620 \text{ kg}^{-1/3} \text{ m}$, ρ_w is the density of liquid water, and Γ_{eff} is the adiabatic rate ($\text{g m}^{-3} \text{ km}^{-1}$) of increase in the liquid water content with height (see also Kubar et al., 2009 for more details of derivation). We note that satellite data inherently include uncertainties stemming from retrieval assumptions, which are not replicated in the model output. Although it could be a part of the reason for discrepancies between the model and observations, this would

mostly be canceled when susceptibilities of cloud and precipitation to aerosol loading are evaluated by a logarithmic form. This study applies the uncertainty thresholds of < 5 and $< 1 \mu\text{m}$ for τ_c and r_e from the 2B-TAU product, respectively, which contributes to the reduction of the retrieval uncertainty of N_c described above as much as possible (Michibata et al., 2014).

To examine the cloud-to-rain conversion process, the conversion rate (P_{conv}) contributed by both autoconversion (collision-coalescence of cloud droplets) and accretion (collision of cloud droplets by raindrops) was derived from the approximation suggested by Stephens and Haynes (2007). This method is established by the continuous collection equation (Pruppacher and Klett, 1997) using observed drop size distributions. P_{conv} was estimated from MODIS LWP and Cloud-Sat mean cloud-layer radar reflectivity $\bar{Z}$ as

$$P_{\text{conv}} = c_1 \text{LWP} \bar{Z} H[Z - Z_c], \quad (4.5)$$

where $c_1 = \kappa_2/2^6$ is a coefficient from the collection kernel (Long, 1974) with $\kappa_2 = 1.9 \times 10^{11} \text{ cm}^{-3} \text{ s}^{-1}$ and sixth moment factor with radar reflectivity. $H[Z - Z_c]$ is the Heaviside step function to exclude the cases that are less than the critical radar threshold Z_c of -15 dBZ_e for which the conversion process is negligible (Matrosov et al., 2004). Although this formulation is based on marine stratocumulus cases from DYCOMS-II measurements (vanZanten et al., 2005), it is applicable for global analysis to study aerosol-cloud interactions (Stephens and Haynes, 2007; Sorooshian et al., 2013) in drizzling light-rain cases ($\bar{Z} < 0 \text{ dBZ}_e$). The parameterization and assumptions used in this method (Eq. 4.5) are also valid for comparison between observations and model simulation (Suzuki and Stephens, 2009). This brings valuable understanding for microphysical conversion processes and its timescales, which matches the scope of our study.

Lower-tropospheric stability (LTS) was derived from the ECMWF-AUX product as the difference in potential temperatures between 700 hPa and the surface (Klein and Hartmann, 1993), and is used for a metric of macroscopic thermodynamic conditions.

4.3 Results

4.3.1 Precipitation Microphysics

A commonly known problem in GCMs associated with low cloud precipitation microphysics is the timing of precipitation and its frequency and/or intensity (Suzuki et al., 2013a, 2015). This issue is also related to the magnitudes of the aerosol indirect effect, i.e., dependency of precipitation

on N_c . As a proxy for this, we use the precipitation susceptibility (S_p) metric, defined as

$$S_p = -\frac{d \ln R}{d \ln N_c}, \quad (4.6)$$

where R is the rain rate. Observed values of R are derived from CloudSat radar reflectivity through the $Z - R$ relationship (Comstock et al., 2004), while model values of R are obtained as the large-scale precipitation rate. We apply a threshold of $R > 0.14 \text{ mm day}^{-1}$, which is equivalent to a radar reflectivity of -15 dBZ_e (Terai et al., 2015), for precipitation in the model to enable a fair comparison with satellite observations. The S_p metric is useful for examining the aerosol impact on precipitation (Sorooshian et al., 2009; Feingold and Siebert, 2009).

Figure 4.1a shows the behavior of S_p as a function of the LWP obtained from MIROC and A-Train satellite observations. The satellite S_p increases with increasing LWP up to around LWP $\sim 450 \text{ g m}^{-2}$; further increases in LWP result in a decrease in S_p . This behavior can be interpreted as follows (Sorooshian et al., 2009). S_p is low for a low LWP because clouds cannot generate much rainwater, regardless of the aerosol loading. At a high LWP, S_p is also low because precipitation is dominant, regardless of the aerosol loading, due to the abundant LWP. In other words, the LWP value at which the S_p peaks corresponds to the turning point where the water conversion process shifts from the autoconversion regime to the accretion regime, as suggested by previous studies (Wood et al., 2009; Sorooshian et al., 2009).

On the other hand, the S_p amplitude predicted by MIROC is smaller than the satellite results for a wide range of LWPs, and S_p remains high even after its values peak near LWP $\sim 450 \text{ g m}^{-2}$. This is mainly because the autoconversion parameterization assumes a constant dependency on N_c (i.e., β) regardless of the LWP, as is clear from Eq. (4.1). This also leads to a significant overestimation of S_p for LWP $< 100 \text{ g m}^{-2}$, which means that the model readily generates rainwater even when only a small amount of cloud water is present.

To understand the uncertainty in the conversion process from cloud water to rainwater in more detail, we define a new metric, the “susceptibility of microphysical conversion (S_{conv})” as

$$S_{\text{conv}} = -\frac{d \ln P_{\text{conv}}}{d \ln N_c}. \quad (4.7)$$

This metric represents how aerosol burdens suppress rainwater production. In satellite analysis, P_{conv} can be estimated from the method proposed by Stephens and Haynes (2007), as shown in Eq. (4.5). In the model, P_{conv} is obtained as a native output of the process rate. The method of Stephens and Haynes (2007) was compared with a native model output of the process rate in a global cloud-resolving model (CRM) by Suzuki and Stephens (2009). The study showed that the

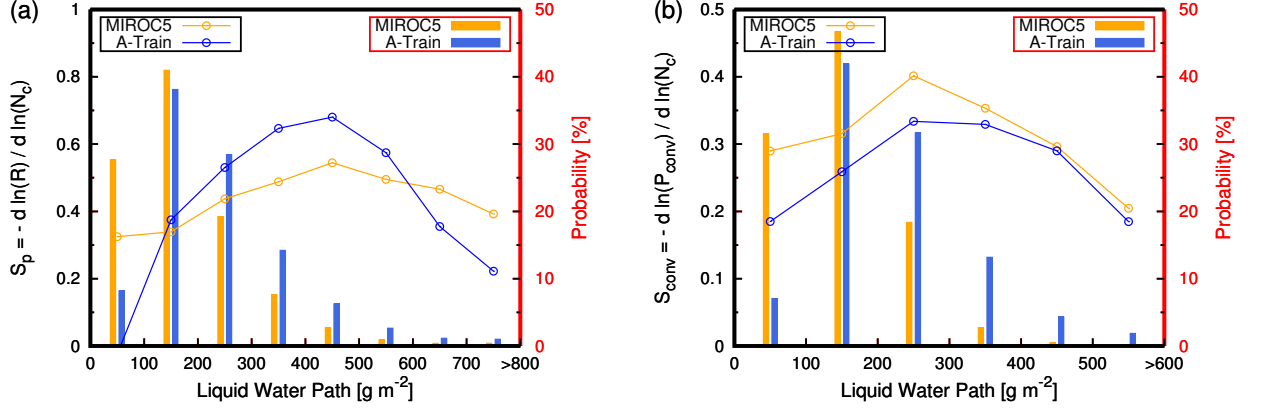


Fig. 4.1: Susceptibilities of (a) precipitation S_p and (b) microphysical conversion S_{conv} as a function of the liquid water path (LWP) for the MIROC-SPRINTARS results and A-Train observations. The left axis shows the value of the susceptibility (refer to the line graph), and the right (red) axis shows the probability distribution function for each LWP bin (refer to the bar chart).

radar reflectivity is a gross measure of the water conversion timescale, supporting the underlying assumption of Stephens and Haynes (2007). This implies that the S_{conv} , which represents the timescale dependency on N_c , can be compared between satellite observations and model simulations although absolute values of P_{conv} can be different between them. Although the use of satellite simulators would be helpful for more direct comparison between model and observations, it is left as the subject of future work.

As shown in Fig. 4.1b, MIROC overestimates S_{conv} , particularly in the lower LWP range. This means that the model generates precipitation at a higher frequency, even at low LWPs, compared to observations, which is mainly because the autoconversion in the model is too rapid (Michibata and Takemura, 2015; Suzuki et al., 2015), as described above. Consequently, the PDF of the LWP is biased toward lower values because cloud water is depleted quickly by the rapid surface precipitation. Alternatively, it is also possible that the model has biases in the condensation processes, which lead to lower LWP and, thus, result in a lower autoconversion rate. These tendencies in the model are strongly related to unrealistically light rain that is too frequent, which is a common problem in GCMs (Stephens et al., 2010), including MIROC.

Besides this, S_{conv} can also be biased from the error of cloud geometric thickness due to insufficient vertical resolution in GCMs. In addition to the microphysical aspects mentioned above, biases in macrophysical structure are also related to model performances, which will be discussed later (see Sect. 3.3).

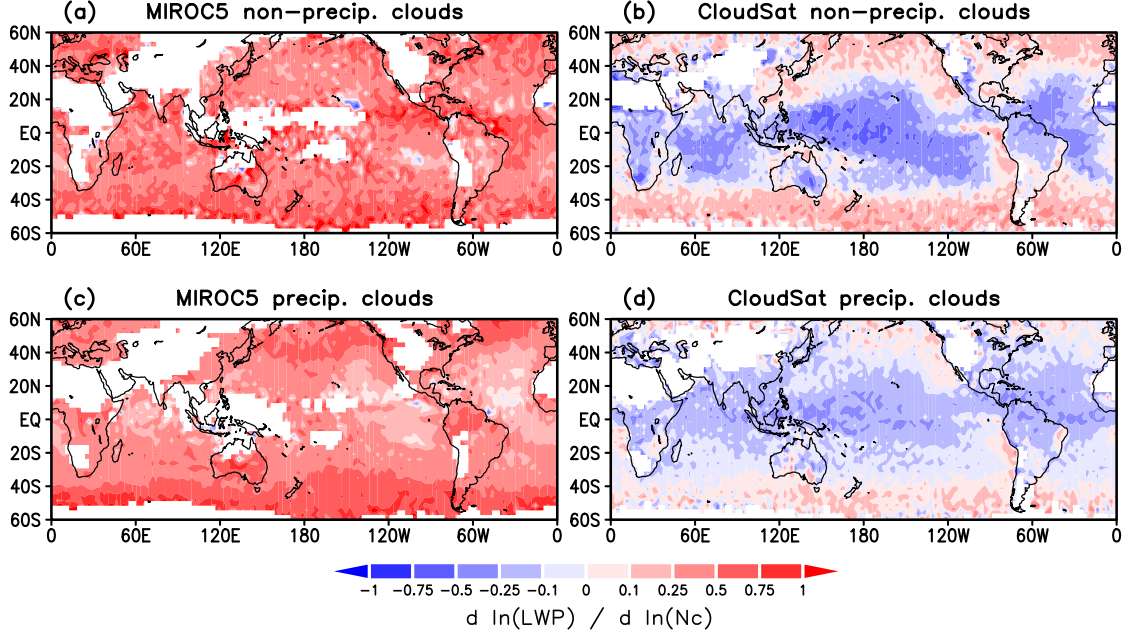


Fig. 4.2: Global distribution of $d \ln LWP / d \ln N_c$ from (a, c) MIROC–SPRINTARS and (b, d) A-Train satellite estimations for non-precipitating and precipitating clouds, respectively. The threshold of the large-scale precipitation rate of 0.14 mm day^{-1} is used to distinguish between non-precipitating or precipitating events (see text for details).

4.3.2 Cloud Susceptibilities

The response of cloud liquid water to aerosol perturbations determines the cloud lifetime via the modification of cloud fraction (Albrecht, 1989), and is thus related to global hydrological cycles as well as radiation budget (e.g., Trenberth et al., 2009; Wood, 2012). As such, it is of great importance to global climate studies to understand why there are two competing mechanisms reported in the literature regarding the pathways of LWP, which cause the LWP to either increase or decrease in response to an increase in aerosols.

Figure 4.2 shows the geographical distributions of LWP susceptibility (i.e., the second term on the right side in Eq. 4.2) obtained from the MIROC simulation and satellite retrievals. The model produces positive values of $d \ln LWP / d \ln N_c$ almost everywhere across the globe in both non-precipitating and precipitating cases, which indicates that the LWP systematically increases with an increasing aerosol burden. Even when the model applies AOD or hygroscopic N_a burden as an aerosol proxy instead of N_c , we obtain the similar results (i.e., globally enhanced LWP). This result is expected from the model parameterization of the cloud lifetime effect (Eq. 4.1), which monotonically delays the onset of precipitation in polluted conditions. This is also a char-

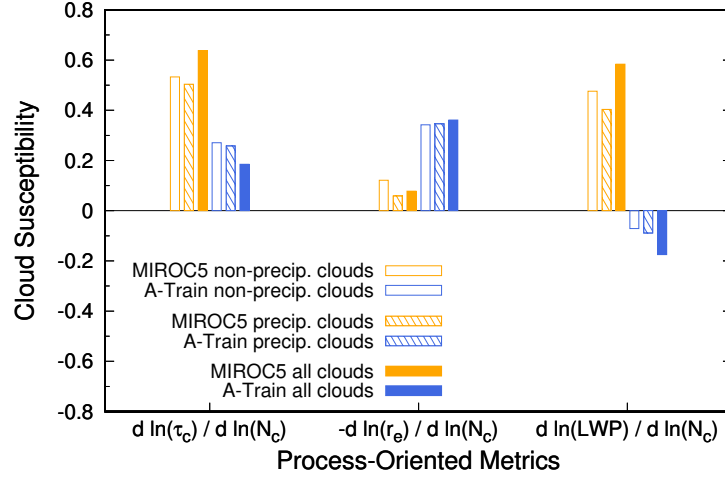


Fig. 4.3: Global mean susceptibility ($60^\circ \text{ S} - 60^\circ \text{ N}$) of τ_c , r_e , and the LWP to N_c . The MIROC result is shown in orange and the A-Train observation is in blue.

acteristic common to other GCMs, as reported in a recent study (Ghan et al., 2016). In contrast, the satellite-derived LWP susceptibility has a coherent geographical pattern that includes both increasing and decreasing responses, which is quite different from the model results. The decreasing response occurs over the tropics and subtropics where more convective cloud is dominant. The increasing response is apparent mainly over the midlatitudes and regions where low clouds are dominant (Klein and Hartmann, 1993).

Nevertheless, it should be noted that Fig. 4.2c captures the horizontal distribution of LWP susceptibility, whose pattern is very similar to observations. That is, the relationship becomes weaker towards the tropics, although the sign is still different. One of the possible mechanisms is the dominance of cloud dynamical processes with high natural variability over tropical and/or subtropical oceans rather than microphysical modifications by aerosols (Peters et al., 2011, 2014). The same processes observed from satellites could be at work in the model, and hence it might be related to the parameterization of subgrid-scale variability. However, this is not always true, particularly in non-precipitating cases (Fig. 4.2a), so we must interpret the mechanisms carefully with further analysis in future.

Figure 4.2b and d show that the geographical patterns are qualitatively similar between the non-precipitating and precipitating conditions, whereas the contrast between the two is slightly different. More specifically, the value of $d \ln LWP / d \ln N_c$ is smaller in the precipitating condition than in the non-precipitating case, which implies a smaller effect of aerosols when precipitation occurs. However, it is noteworthy that the positive response of $d \ln LWP / d \ln N_c$ in the non-

precipitating condition has a negative value in the precipitating conditions over East Asia, the eastern United States, and Europe, where the anthropogenic aerosol burden is severe. This suggests that aerosols act to prolong the cloud lifetime in non-precipitating conditions, while they enhance cloud evaporation or can be a precipitation driver in precipitating conditions which ultimately result in less cloud water. These two competing mechanisms are reasonably consistent with theories suggested by recent studies (Stevens and Feingold, 2009; Rosenfeld et al., 2014; Lebo and Feingold, 2014), which propose the existence of a buffering effect in the cloud system that results in smaller-magnitude aerosol–cloud interactions. These comparisons suggest that the model does not appropriately represent the buffering effect which compensates for the positive responses of the LWP to aerosol perturbations. Current GCMs which use similar parameterization frameworks (e.g., autoconversion) therefore inherently overestimate the aerosol indirect effect as reported by previous studies (Quaas et al., 2009; Wang et al., 2012; Gettelman et al., 2013).

Figure 4.3 summarizes the relationship among the “process-oriented metrics” corresponding to each term in Eq. (4.2). The global mean susceptibilities (averaged from 60° S to 60° N) for each term were calculated from the individual susceptibility in each grid box in which more than 10 warm cloud samples were obtained, which contributed to the reduction of statistical noise.

The cloud susceptibility of τ_c to N_c in MIROC is approximately twice as large as in the A-Train results. The significant bias is decomposed into contributions due to the “albedo effect” $-d \ln r_e / d \ln N_c$ and the “lifetime effect” $d \ln \text{LWP} / d \ln N_c$. Although the former is underestimated in the model compared with A-Train estimations, its sign is positive and consistent with observations. The overestimation of τ_c -susceptibility in the model is therefore attributed to a positive response of the LWP, which is in stark contrast to the slight negative responses in satellite observations. Ghan et al. (2016) reported a wide diversity in the relationship between the LWP and N_c among nine AeroCom GCMs, with all models showing an enhanced response of LWP to increased N_c . This further causes uncertainties in the estimation of radiative forcing (Ghan et al., 2016; Feingold et al., 2016).

As Figs. 4.2 and 4.3 indicate, the discrepancy in the LWP response between the model and observations can be a critical source of model uncertainty, which causes a bias in climate responses via the aerosol–cloud–precipitation–climate interaction.

4.3.3 Dependency of LWP Responses on Meteorology

Another key question regarding the LWP response for aerosol perturbations is how and to what extent it depends on macrophysics such as cloud regimes and thermodynamic conditions in the real atmosphere. Recent studies have suggested that the source of uncertainty in the LWP response could be attributed to differences in meteorology, different cloud types and regimes, or more theoretical reasons, based on satellite observations (Sorooshian et al., 2013; Chen et al., 2014), large-eddy simulation (LES; Lebo and Feingold, 2014; Seifert et al., 2015), and GCM intercomparison (Wang et al., 2012; Zhang et al., 2016).

To address this question, we examine the dependency of the LWP susceptibility on both column maximum radar reflectivity ($Z_{\max}$) and LTS as shown in Fig. 4.4. Given that the horizontal axis characterizes the rain regime (i.e., non-precipitating, drizzling, or precipitating) and the vertical axis represents the thermodynamical stability conditions (i.e., unstable, intermediate, or stable), the diagram illustrates how the LWP response to perturbed aerosols varies as a function of both rain characteristics and stability conditions, thus providing a way to classify cloud susceptibility according to macroscopic and meteorological conditions.

Figure 4.4 clearly shows a systematic variation of the cloud susceptibility under the two conditions. Positive responses of the LWP to N_c are dominant in the non-precipitating and stable environments, while negative responses can be seen in precipitating and unstable conditions. The top-left region in the diagram corresponds to a stratocumulus regime in the marine boundary layer. Because this type of cloud typically produces light precipitation (i.e., drizzle; Wood, 2012) rather than heavy precipitation it depletes a large amount of cloud water, and the aerosols ingested into this type of cloud effectively act to enhance cloud water storage, resulting in a positive response of LWP to an increased aerosol loading. In contrast, the right-bottom region in the diagram corresponds to a more convective cumulus regime, which is present mainly over the tropics. This type of cloud is characterized by a relatively fast precipitation timescale (Sorooshian et al., 2013), and is favorable for cloud water evaporation due to the larger extent of entrainment mixing (Small et al., 2009), which results in negative responses of the LWP.

It is interesting that there is a positive correlation in the top-right region even though precipitation occurs. One possible interpretation of this is that the water-vapor supply is dominant over the loss of cloud water by precipitation, and this type of cloud may correspond to the sustained frontal precipitation (precipitating nimbostratus) systems found mainly over midlatitude oceanic regions, where water vapor is abundant. In pristine and clean environments, referred

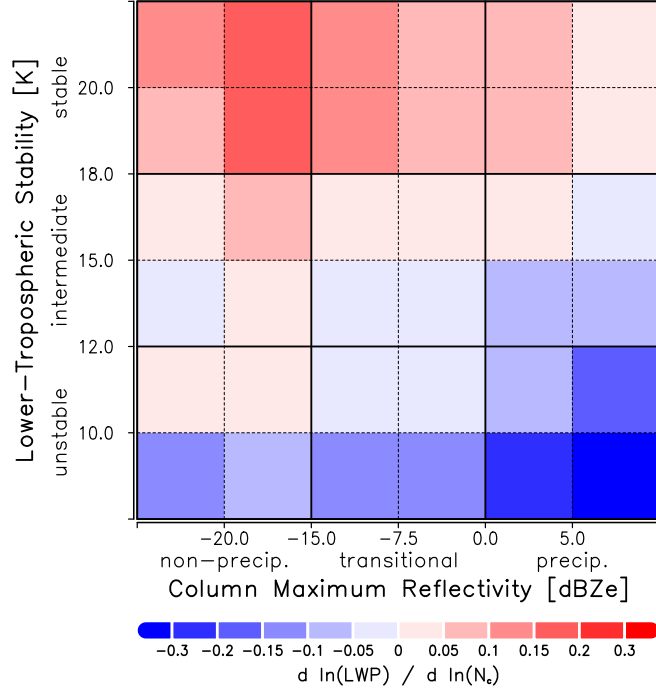


Fig. 4.4: Susceptibility matrix of the LWP response to N_c as a function of column maximum radar reflectivity ($Z_{\max}$) and lower-tropospheric stability (LTS) based on A-Train satellite data.

to as “aerosol-limited” conditions (Koren et al., 2014), aerosols ingested into clouds will tend to store the cloud water but also simultaneously produce more rain due to abundant water mass. We note that it is just a speculation at this stage, and it might be related to background aerosol number and environmental conditions (cf. Sect. 4 for more discussion). It is also noteworthy that the bottom-left region displays negative susceptibilities even though precipitation does not occur. Non-precipitating clouds in a significantly unstable environment would correspond to inland and/or daytime cumulus. This tendency agrees with the results of a previous study (Small et al., 2009) that focused on the non-precipitating cumulus regime, and suggests a mechanism whereby the LWP decreases with increased aerosol loading via evaporation–entrainment feedback. This results in a loss of cloud water without precipitation.

Although the model version of the LWP-susceptibility diagram is not shown, it will indicate a positive value in the matrix overall, as is obvious from Fig. 4.2. The mechanisms proposed above must be confirmed by more detailed examinations using GCM and CRM with satellite simulators, or using high-resolution process modeling, such as LES, in future studies. However, the observation-based findings described above strongly suggest that rigorous studies focusing on macrophysical conditions, including regional characteristics of meteorological factors, in addi-

tion to microphysical conditions, are indispensable for better understanding the response of the aerosol–cloud–precipitation interaction.

4.4 Summary Perspectives

We explored the source of discrepancy in the aerosol–cloud–precipitation interaction for warm clouds between an aerosol–climate model and A-Train satellite retrieval. The instantaneous model output was analyzed using as many samples as possible to provide reliable statistics and fair comparisons with satellite observations.

We found critical biases in the model in the response of the LWP to aerosol perturbations. The model predicted a monotonic increase in the LWP across the globe, in contrast to the observations that clearly showed a regional variation of the LWP response that either increased or decreased with an increasing aerosol loading. This variability in cloud susceptibility observed by the A-Train was closely related to differences in meteorological factors, such as cloud regimes and thermodynamic conditions. For example, stratiform clouds under stable conditions had a tendency to increase the LWP given aerosol perturbations, while cumulus clouds over an unstable environment tended to decrease the LWP as the aerosol loading increased. The bidirectional responses of LWP (both positive and negative) found in satellite observations in different aerosol concentrations might be related to the concept of “optimal aerosol concentration (N_{op})”, recently suggested by Dagan et al. (2015a,b). More specifically, in case of $N_a < N_{\text{op}}$, clouds tend to be deeper with larger liquid mass, referred to as cloud invigoration (e.g., Koren et al., 2014), for increased aerosol loading, whereas the case of $N_a > N_{\text{op}}$ would be favorable for cloud suppression due to enhanced entrainment and evaporation. This could lead the bidirectional LWP susceptibilities, although we cannot mention the exact mechanisms at this stage because N_{op} also depends on both cloud geometric scale and environmental conditions (Koren et al., 2014; Dagan et al., 2015a,b) as well as aerosol types.

This can explain why previous studies have reported conflicting results for the LWP response, with either an increase or decrease with increasing aerosol loading (e.g., Sekiguchi et al., 2003; Ackerman et al., 2004; Matsui et al., 2006). Previous studies have focused on different study regions and/or targets, which has resulted in different cloud responses due to the different mechanisms of aerosol–cloud–precipitation interaction. This means that global-mean cloud susceptibility is not very meaningful in constraining the aerosol–cloud–precipitation interaction. Future studies should consider the regional dependence of the susceptibility metrics (Terai et al., 2015).

The monotonic increase in the LWP with increasing aerosol loading in the model is attributed to the autoconversion scheme, which assumes only suppression of rainwater generation to account for the traditional cloud lifetime effect without its compensation, and does not take meteorological conditions into account. Mechanisms that can decrease the LWP in polluted conditions, such as the enhancement of evaporation due to entrainment mixing (Small et al., 2009; Seifert et al., 2015), are not incorporated in our model. This means that the model fails to represent the buffering effect for the aerosol–cloud–precipitation interaction (Stevens and Feingold, 2009). Moreover, the model overestimates S_{conv} around low LWPs compared with A-Train satellite retrievals due to uncertainties in process rate parameterization (Wood, 2005a). This is evidence that the autoconversion in the model is too fast, which results in the LWP having a high dependency on the N_c . This bias in the model is consistent with a previous study that reported a higher LWP susceptibility in GCMs due to their diagnostic treatment of rainwater (Wang et al., 2012).

In future studies, the aerosol–cloud–precipitation framework must be expanded to represent the effect of environmental conditions in a flexible manner, in addition to the microphysics. Current microphysical frameworks without such macrophysical aspects bring highly sensitive aerosol–cloud interactions, although they vary to some extent depending on the autoconversion scheme (Gettelman, 2015; Suzuki et al., 2015; Michibata and Takemura, 2015). Ghan et al. (2016) also reported a wide diversity in the LWP response to N_c among various GCMs, and concluded that their inconsistency could mainly be attributed to their different representations of the autoconversion process. However, it is also true that different choices of microphysical scheme alone do not significantly improve the model biases in both cloud physics and cloud radiative effects (Michibata and Takemura, 2015), and these two requirements sometimes contradict each other (Suzuki et al., 2013b). This is due to the arbitrary nature of tuning and assumptions (e.g., artificial threshold parameters, diagnostic treatment of rain), which is a bottleneck in GCMs (Hoose et al., 2009; Quaas et al., 2009; Ghan et al., 2013). Recently, a fundamental model improvement was achieved by introducing a prognostic precipitation framework (Gettelman et al., 2015; Sant et al., 2015), which represents important progress in process representations for more realistic cloud and precipitation microphysics. This improvement is expected to overcome some of the common problems in GCMs, such as the overestimation of the aerosol indirect effect and spurious light rain (Walters et al., 2014; Gettelman et al., 2015; Sant et al., 2015).

Furthermore, a representation of subgrid-scale fluctuations has also been critical problem in GCMs. Although the magnitude as well as sign of LWP susceptibility differs between the model

and observations, the horizontal pattern is similar in precipitating conditions. The parameterization of subgrid-scale variability may partly contribute to the weakening of the aerosol roles by capturing the large natural variability of clouds, especially over tropical and/or subtropical oceans (Peters et al., 2011, 2014), which would lead to a more realistic representation of cloud dynamical processes. For example, Guo et al. (2011, 2015) showed that both positive and negative LWP responses can be represented in even a GCM framework, by the PDF-based macrophysics parameterization, called “Cloud Layers Unified By Binormals (CLUBB; Larson and Golaz, 2005)”. Lebsock et al. (2013) estimated a weighting factor of process rate equations to consider the subgrid effects based on A-Train retrievals unless the accretion process is significantly underestimated. The interaction between microphysics and subgrid-scale dynamics (microphysics–dynamics interactions) in GCMs is therefore one of the indispensable processes for incorporating buffering effects and for improving model physics as a whole.

Although this study focused only on warm-phase clouds, our findings regarding different cloud responses to aerosol perturbations between GCMs and satellite observations will assist future model development for more accurate climate simulations. Further studies should also contain an extension of the research target from liquid to mixed and/or iced clouds, and from a process-level to a cloud system to understand the whole cloud system response to aerosol perturbations, taking into account the buffered system morphology.

5 Development of A Physically Based Two-Moment Bulk Microphysics Scheme with Prognostic Precipitation

Abstract

Maritime stratocumuli play a significant role in forming climate because of their optical brightness having strong radiative cooling effects. They frequently form drizzles, whose micro- and macro-physics are highly susceptible to the aerosol forcing as well as boundary layer physics, hence most of global climate models (GCMs) suffer from the representation of stratocumulus and its radiative effects. The diagnostic treatment of rain (DIAG) in GCMs is therefore a bottleneck to reproduce the complex aerosol-cloud-precipitation interactions. Here the prognostic equations both for mass and number mixing ratios of rainwater are introduced into the MIROC-SPRINTARS aerosol-climate model. The new scheme (PROG) is firstly evaluated in a simple zero-dimensional box model and a single-column model. The sensitivity experiments provide that the cloud-to-rain conversion process requires approximately 60 sec, and the use of a modified Gamma function for the raindrop size distribution is favorable for simulating stratocumuli rather than an exponential size distribution. Then the two-moment based PROG scheme is implemented into MIROC-SPRINTARS. As expected, relative contribution of accretion against autoconversion increases significantly since the prognostic treatment of rain can memory the rainwater across multiple time steps. The PROG simulation represents the coastal stratocumuli in good agreement with satellite and in-situ measurements. This further leads to improve the shortwave cloud radiative forcing that has been significantly overestimated in the traditional DIAG scheme. These findings firmly promise the importance of the prognostic precipitation framework, which will overcome the common problems of GCMs in aerosol-cloud-precipitation-climate interactions.

Keywords: prognostic rain scheme, cloud-to-rain conversion, drizzle, stratocumulus, radiation

Key Points:

- The prognostic equations for mass and number mixing ratios are newly introduced in GCM.
- The performance of the new scheme is evaluated in both a single-column model and global model simulations.

- Drizzling coastal stratocumuli are well reproduced in the new scheme, which improves shortwave cloud radiative forcing.
-

5.1 Introduction

Stratocumulus covers more than 20% of the Earth’s area, and has a strong influence on climate systems via its cooling contribution (Wood, 2012). The boundary layer low-clouds particularly interact with aerosols, referred to as “aerosol-cloud interactions (ACI)” (Twomey, 1977; Albrecht, 1989), but also correlate to environment and thermodynamic conditions (Klein and Hartmann, 1993; Wood and Hartmann, 2006), whose complex morphology plagues emulation of cloud micro- and macrophysics (e.g., Wang et al., 2012; Komurcu et al., 2014; Calisto et al., 2014) as well as cloud system responses to aerosol perturbations (Stevens and Feingold, 2009; Michibata et al., 2016; Zhou and Penner, 2017) in general circulation models (GCMs).

GCMs tend to overestimate the shortwave cloud radiative forcing (SWCRF) even though the low-cloud fraction is underestimated (Nam et al., 2012). This “too few too bright low-cloud problem” is a common issue in the practically all of GCMs (Calisto et al., 2014). Nam et al. (2012) argued that the importance in modeled parameterization of boundary layer processes since they found the biases in the dependence of cloud vertical structures on environmental regimes. This issue closely relates to the magnitude of the ACI that is overestimated in GCMs (e.g., Quaas et al., 2009; Wang et al., 2012). Zhang et al. (2016) found large spreads among GCMs in the response of liquid water path (LWP) to aerosol perturbations for different dynamic regimes, which implies GCMs do not capture the regime dependence of the aerosol on microphysics-dynamics interactions (Wood, 2012; Michibata et al., 2016).

The uncertainty in the sensitivity of cloud lifetime to aerosols is inevitably linking to the uncertainty in the rain formation. The cloud-to-rain hydrological conversion in models is represented by the so-called autoconversion process. The autoconversion rate is generally parameterized as a function of the cloud liquid water mixing ratio q_c and the cloud droplet number concentration N_c (e.g., Berry, 1968; Khairoutdinov and Kogan, 2000; Beheng, 1994) as,

$$\left. \frac{\partial q_r}{\partial t} \right|_{aut} = C_{aut} q_c^\alpha N_c^{-\beta} \quad (5.1)$$

where q_r is the rainwater mixing ratio, and C_{aut} , α , and β are the prescribed constants. The increased N_c due to increased aerosols thus decreases the autoconversion rate to prolong the

cloud lifetime. This is why the cloud-to-rain microphysical transition is highly sensitive to the autoconversion scheme used (Wood, 2005a; Suzuki et al., 2013a).

Suzuki et al. (2015) evaluated the rain formation process through a multi-model analysis (both GCMs and a cloud resolving model; CRM) using satellite observations. GCMs represent rain formation too early than the satellite estimate, even when the drizzle initiation rarely occurs at smaller cloud droplet size $< 10 \mu\text{m}$ (e.g., Nakajima et al., 2010; Suzuki et al., 2011; Michibata et al., 2014). On the other hand, only the CRM characterized by prognostic treatment of rain (hereinafter “PROG”) could capture cloud-to-rain transition process depicted by the satellite data, which suggests that diagnostic treatment of rain (hereinafter “DIAG”) assumed in most GCMs may be missing the intrinsic cloud-precipitation microphysics. For example, although shallow raindrops (drizzle) which do not reach to the surface within a single model time step exist in the real atmosphere, the DIAG method assumes the all of the diagnosed rainwater precipitate to the surface within the time step. This ad hoc treatments would further cause biases in precipitation frequency and/or intensity (Stephens et al., 2010; Suzuki et al., 2015) as well as radiative properties (Muhlbauer et al., 2014; Michibata and Takemura, 2015).

It should be noted that there is a critical discrepancy between “process-oriented constraints” of microphysics estimated from satellite (Suzuki et al., 2013b) and “performance-oriented constraints” of the climate trend with an empirical model tuning (Golaz et al., 2013). In other words, this can be interpreted as a signal of the error compensation in modeled parameterization and historical temperature trend via the arbitrary “tunable knobs” in GCMs (Suzuki et al., 2013b, 2015). These studies strongly claim that the modeling framework of the aerosol-cloud-precipitation interactions must be elaborated at a fundamental process-level (Seinfeld et al., 2016) toward a more physically based parameterization.

Recently, advanced framework for cloud-precipitation microphysics was introduced with a prognostic treatment of rain into GCMs (Gettelman et al., 2015; Sant et al., 2015). They reported that the ratio of accretion to autoconversion (Acc/Aut ratio) was significantly improved due to enhancement of the accretion process (Posselt and Lohmann, 2008), which is reasonable for in-situ measurements (Wood et al., 2011; Wood, 2005a; Gettelman et al., 2013). The enhancement of the relative contribution of accretion over autoconversion results in relaxation of the aerosol forcing (Posselt and Lohmann, 2009; Gettelman et al., 2013) since the aerosol effect depends only on autoconversion process via its relation with N_c as described in Eq. (5.1) whereas accretion process is independent from aerosol (or cloud droplet) number.

The main objective of this Chapter is to clarify how different treatment of rain (i.e., DIAG vs. PROG) in GCMs will contribute to the relevant processes of the aerosol-cloud-precipitation-climate interactions. To this end, this Chapter develops the new cloud-precipitation framework with PROG scheme, and implements it into MIROC-SPRINTARS GCM. The performance and advantages of the new PROG scheme against to the DIAG approach will be evaluated in both single-column model (SCM) and global model using satellite observations.

5.2 Description of the Scheme and Model

In the traditional DIAG method, the diagnosed precipitating hydrometeors (i.e., rainwater and snow water) at a present time step fall out to the surface instantaneously (Ghan and Easter, 1992). This approach is reasonable for relatively large hydrometeors with their fall velocities of $3\text{--}5\text{ ms}^{-1}$, which will fall 1800–6000 m in the typical time step (i.e., 10–20 min) applied in GCMs. The DIAG approach is not constrained by the Courant–Fredrichs–Levy (CFL) condition in sedimentation process of precipitating hydrometeors, which is convenient for GCMs with long time step to reduce the computational cost. Alternatively, the DIAG method inevitably neglects small raindrops or snowflake which reside longer time scale than model time step. This would further cause biases in the hydrological cycle and/or vertical distribution of hydrometeors since the DIAG omits evaporation (sublimation) during sedimentation of drizzle (snowflake) as well as thermal transport due to their processes.

Here the existing DIAG scheme in MIROC-SPRINTARS has been reconstructed significantly as shown in Fig. 5.1. The new scheme includes prognostic precipitation based on two-moment approach, namely, it solves prognostic equations for both rain mass (q_r) and number (N_r) mixing ratios. The time tendencies of them are described as,

$$\frac{\partial q_r}{\partial t} = -\frac{1}{\rho_a} \nabla \cdot (\rho_a \mathbf{u} q_r) - \frac{1}{\rho_a} \frac{\partial(\rho_a q_r v_q)}{\partial z} + S_{q_r} \quad (5.2)$$

$$\frac{\partial N_r}{\partial t} = -\frac{1}{\rho_a} \nabla \cdot (\rho_a \mathbf{u} N_r) - \frac{1}{\rho_a} \frac{\partial(\rho_a N_r v_N)}{\partial z} + S_{N_r} \quad (5.3)$$

where ρ_a is the air density and $\mathbf{u}$ is the wind vector. The v_q and v_N are the bulk fall velocities weighted by mass and number, respectively, and described later (cf. Eqs. 5.7 and 5.8). The S_{q_r} and S_{N_r} are the source/sink of mass and number mixing ratios with respect to condensation, evaporation, and other microphysical collection processes (e.g., autoconversion, accretion, and self-collection). These “conservation equations” were newly introduced into MIROC-SPRINTARS GCM, and now aerosol-cloud micro- and macrophysics link with precipitation interactively. Note

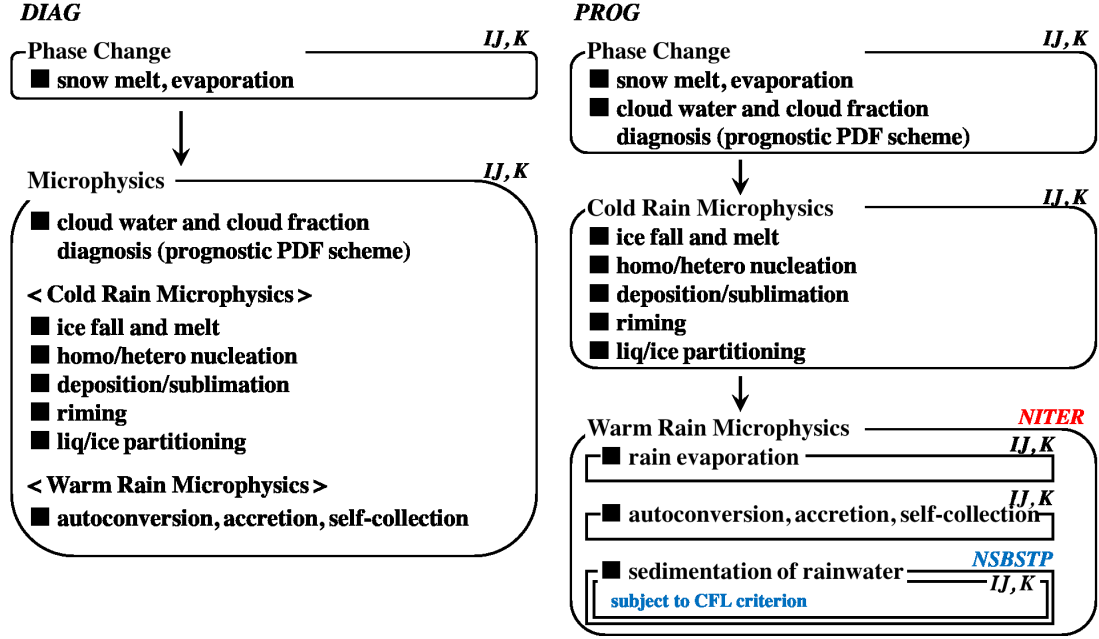


Fig. 5.1: Schematic summary of cloud physics calculation for DIAG and PROG schemes. The “*IJ*” and “*K*” at the top right of the box indicate horizontal and vertical loop, respectively. The “*NITER*” and “*NSBSTP*” mean sub-time step iteration with respect to microphysics and sedimentation, respectively (see text for details).

that the solid precipitation is still a diagnostic variable at the current version since this Chapter primarily aims to improve warm rain microphysics and their radiative properties.

The raindrop size distribution $N(D)$ is based on the modified gamma distribution (Ulbrich, 1983), which is described as,

$$N(D) = \frac{N_r}{\Gamma(\mu)} \left(\frac{D}{D_0} \right)^\mu \frac{1}{D} \exp \left(-\frac{D}{D_0} \right) \quad (5.4)$$

where $\Gamma(\mu)$ is the Gamma function, and N_r and D are the total raindrop number concentration [m^{-3}] and the raindrop diameter [m], respectively. $N(D)$ depends upon both the shape parameter μ and the distribution parameter D_0 described as,

$$D_0 = \frac{\overline{D}}{\sqrt[3]{(\mu+2)(\mu+1)\mu}} \quad (5.5)$$

where the mean diameter of raindrop $\overline{D}$ [m] is,

$$\overline{D} = \sqrt[3]{\frac{6}{\pi \rho_w} \frac{\rho_a q_r}{N_r}} \quad (5.6)$$

ρ_w is the density of water assuming 1000 [kg m^{-3}].

The mass and number weighted bulk fall velocities v_q and v_N , which are the approximation based on Rogers et al. (1993), are described as follows (Posselt and Lohmann, 2009; Sant et al.,

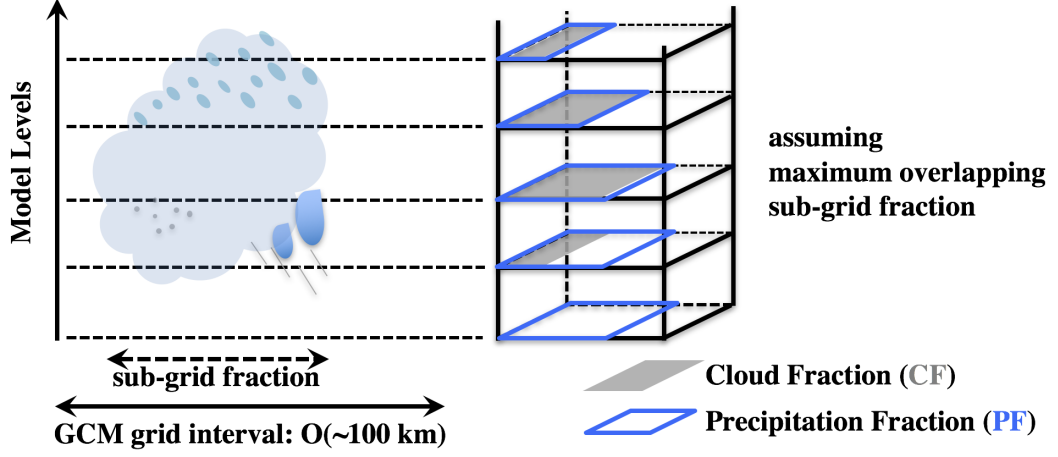


Fig. 5.2: Schematic of the vertical overlap method for the precipitating hydrometer. Precipitation is distributed by assuming only the part of maximally overlapping clouds or precipitation above the layer (see text for details).

2015):

$$v_q = \begin{cases} (\mu + 3) b_v D_0 & \text{for } D_0 < D_v/(\mu + 3) \\ b_1 & \text{for } D_0 > D_v/(\mu + 3) \end{cases} \quad (5.7)$$

$$v_N = \begin{cases} \mu b_v D_0 & \text{for } D_0 < D_v/\mu \\ b_1 & \text{for } D_0 > D_v/\mu \end{cases} \quad (5.8)$$

with the constants $b_1 = 9.65 \text{ [m s}^{-1}\text{]}$, $b_v = 3918 \text{ [s}^{-1}\text{]}$, and $D_v = 2463 \text{ [}\mu\text{m}]$. Thus the sedimentation of the mixing ratios in flux forms are expressed as,

$$\left. \frac{\partial q_r}{\partial t} \right|_{sed} = \frac{1}{\rho_a} \frac{\partial}{\partial z} (\rho_a q_r v_q) \quad (5.9)$$

$$\left. \frac{\partial N_r}{\partial t} \right|_{sed} = \frac{1}{\rho_a} \frac{\partial}{\partial z} (\rho_a N_r v_N) \quad (5.10)$$

which are strictly constrained by the vertical CFL criteria so that $C = v_q \Delta t / \Delta z < 1$ will be satisfied, as shown in Fig. 5.1 as “NSBSTP”.

The subgrid fraction of precipitation is considered by assuming the maximum overlap as shown in Fig. 5.2. The precipitation fraction is diagnosed based on the cloud fraction derived from the PDF-based prognostic large-scale condensation (LSC) scheme (Watanabe et al., 2009). The both cloud liquid and ice are prognostic variables based on a two-moment treatment (Takemura et al., 2009). Solid precipitation is treated as a diagnostic manner in this study based on Wilson and Ballard (1999) microphysics scheme, and calculated within the standard model time step (12 min in this study) whereas liquid precipitation is treated within the sub-time step as described earlier.

The sensitivities in the prescribed parameters and microphysics scheme such the shape parameter in raindrop size distribution (DSD), cloud-to-rain conversion scheme, and length of the sub-time step are discussed in the following section.

5.3 Sensitivity Experiments in Idealized Cases

The prognostic variables for rainwater (i.e., q_r and N_r) will contribute more physically based hydrological interactions among water vapor, cloud liquid water, and cloud ice explicitly. The implementation of the PROG scheme into GCM involves the introduction of the new microphysical processes (e.g., accretion, self-collection, and sedimentation). To understand the sophisticated interactions among hydrometeor classes, first we apply the PROG scheme to the simple model framework for process-level verification. The results from zero dimensional box model (cf. Section 5.3.1) and from SCM (cf. Section 5.3.2) are presented below.

5.3.1 Box Model Tests

The tendency terms in microphysical processes significantly control the hydrological cycle among the hydrometer classes. Here we prepare a simple zero-dimensional box model which calculates mass and number tendencies of each microphysical process for prescribed initial parameters (e.g., temperature, height, water vapor, mass and number of hydrometeors). The processes considered in the box model are summarized in Table 5.1. Note that the model includes processes with respect to snow as well. The source and sink terms of cloud water (i.e., condensation and evaporation) are not shown because this test focuses only on interactions with different hydrometeor classes and collection processes. The references in Table 5.1 are the standard scheme implemented into the new PROG scheme, and its sensitivity on model configurations will be discussed later (cf. Sections 5.3.2 and 5.4). The code of the box model is provided in Appendix A.

The simulations were carried out for 10 cases of $T = \{266, 267, \dots, 275\}$ K and nine sets of three cases of hydrometeors as listed in Table 5.2, resulting a grand total of 196,830 ($= 10 \times 3^9$) simulations. These parameters were perturbed in realistic values which lie within the range of the in-situ observations (e.g., Wood, 2005b; vanZanten et al., 2005; Wang et al., 2009), satellite estimates (e.g., Lebsock et al., 2011; Li et al., 2012; Zeng et al., 2014), and simulations in GCMs (e.g., Waliser et al., 2011; Jiang et al., 2012; Komurcu et al., 2014).

The mean and standard deviation of each tendency are summarized in Table 5.3, and mass tendencies are shown in Fig. 5.3. It is obvious that the contributions of accretion process in

Table 5.1: Summary of interaction processes among hydrometeors (water vapor “v”, liquid cloud “c”, cloud ice “i”, rain “r”, and snow “s”) for mass (Q) and number (N) implemented into the new prognostic precipitation framework. The check-mark in the left column denotes that the tendency term is incorporated into the global model as well. The up (down) arrows indicate that the process leads to increase (decrease) with regard to the mass and number mixing ratios.

Process	Interaction	Q_c	Q_i	Q_r	Q_s	N_c	N_i	N_r	N_s	References
Prognostic Rain										
✓ autoconversion	$c + c \rightarrow r$	↓		↑		↓		↑		Khairoutdinov and Kogan (2000)
✓ accretion	$c + r \rightarrow r$	↓		↑		↓				Khairoutdinov and Kogan (2000)
✓ self-collection	$r + r \rightarrow r$							↓		Beheng (1994)
✓ evaporation	$r \rightarrow v$			↓				↓		Wilson and Ballard (1999)
✓ sedimentation	r			↑↓				↑↓		Rogers et al. (1993)
brake-up	$r \rightarrow r$							↑		Seifert and Beheng (2001)
Prognostic Snow										
✓ instantaneous melt	$s \rightarrow r$			↑	↓			↑	↓	
✓ freezing of rain	$r \rightarrow s$			↓	↑			↓	↑	
✓ aggregation	$i + i \rightarrow s$		↓		↑		↓		↑	Ferrier (1994)
✓ accretion	$c + s \rightarrow s$	↓			↑	↓				Seifert and Beheng (2006)
✓	$i + s \rightarrow s$		↓		↑		↓			Seifert and Beheng (2006)
✓	$r + s \rightarrow s$			↓	↑			↓		Reisner et al. (1998)
✓ Hallet-Mossop	$c \rightarrow i$	↓	↑				↑			Cotton et al. (1986)
✓ self-collection	$s \rightarrow s$								↓	Passarelli (1978)
sublimation	$s \rightarrow v$				↓			↓		Wilson and Ballard (1999)
✓ sedimentation	s				↑↓			↑↓		Barthazy and Schefold (2006)
✓ Bergeron	$v \rightarrow i$	↓	↑							Rotstayn et al. (2000)
✓	$i + v \rightarrow s$	↓			↑					Nogherotto et al. (2016)
brake-up	$s \rightarrow s$							↑		Seifert and Beheng (2006)

warm rain and Wegener–Bergeron–Findeisen process (Wegener, 1911; Bergeron, 1935; Findeisen, 1938) in cold rain microphysics are dominant, which is consistent with observations (Wood et al., 2011; Cesana et al., 2015) and modeling studies (Gettelman et al., 2010, 2015). These interactive coupling with hydrometeor species in the PROG framework will contribute to more realistic hydrological cycle and distribution as discussed later.

5.3.2 Single-Column Model Tests

Next, the sensitivity of the PROG scheme to model configurations is examined. For instance, the prescribed DSD is required in bulk representation applied here, which has a high impact on rain formation process (e.g., Posselt and Lohmann, 2009; Abel and Boutle, 2012). The use of autoconversion scheme is also sensitive to the cloud-to-rain conversion process (e.g., Wood, 2005a; Michibata and Takemura, 2015; Suzuki et al., 2015). Furthermore, there is poor consensus on the length of the sub-time step iteration in the state-of-the-art cloud-precipitation scheme (Gettelman and Morrison, 2015; Sant et al., 2015), which is especially important in cold rain microphysics

Table 5.2: Prescribed parameters for temperature T and hydrometeors for water vapor, liquid cloud, cloud ice, rain, and snow with regard to mass and number mixing ratios in the box model test.

Parameters	Prescribed values
T (K)	266, 267, 268, 269, 270, 271, 272, 273, 274, 275
Q_v (kg kg^{-1})	8.0×10^{-5} , 5.0×10^{-4} , 1.0×10^{-3}
Q_c (kg kg^{-1})	1.0×10^{-5} , 6.0×10^{-5} , 3.0×10^{-4}
N_c (m^{-3})	3.0×10^7 , 1.0×10^8 , 3.0×10^8
Q_i (kg kg^{-1})	2.0×10^{-6} , 8.0×10^{-6} , 3.0×10^{-5}
N_i (m^{-3})	5.0×10^5 , 1.0×10^6 , 3.0×10^6
Q_r (kg kg^{-1})	6.0×10^{-6} , 4.0×10^{-5} , 8.0×10^{-5}
N_r (m^{-3})	5.0×10^3 , 1.0×10^4 , 3.0×10^4
Q_s (kg kg^{-1})	1.0×10^{-6} , 5.0×10^{-6} , 3.0×10^{-5}
N_s (m^{-3})	5.0×10^3 , 1.0×10^4 , 3.0×10^4

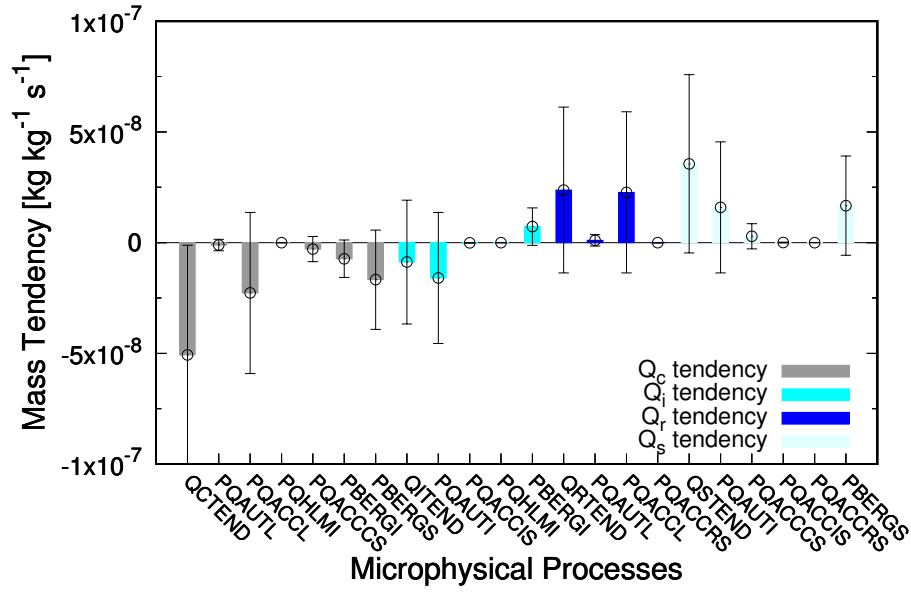


Fig. 5.3: Box model calculations for mass tendencies via interactions described in Table 5.1 with regard to liquid cloud water (grey), cloud ice (cyan), rainwater (blue), and snow (light cyan). The error bars show the standard deviation due to perturbations in parameters described in Table 5.2. The abbreviations with respect to each process are denoted in Table 5.3. The mass source (sink) of cloud water via condensation (evaporation) is not listed (see text for details).

(Gettelman et al., 2015). Such the model configurations should be evaluated in a simple model framework before the scheme is incorporated into a complicated global model.

Here the several sensitivity experiments related to the DSD, autoconversion schemes, and sub-time step are designed as summarized in Table 5.4. The experiments are carried out with the SCM which applies the same scheme in a global version of MIROC. The MIROC-SCM has participated in CGILS (CFMIP-GASS Intercomparison of Large-Eddy Simulations and SCM)

Table 5.3: Box model statistics for tendencies of mass ($\text{kg kg}^{-1} \text{s}^{-1}$) and number ($\text{m}^{-3} \text{s}^{-1}$) in each process described in Table 5.1.

Process	Variation	Abbreviation	Mean (Standard deviation)
autoconversion of cloud water	$Q_c \downarrow$	PQAUTLC	-1.07×10^{-9} (2.60×10^{-9})
	$Q_r \uparrow$	PQAUTLC	$+1.07 \times 10^{-9}$ (2.60×10^{-9})
	$N_c \downarrow$	PNAUTLC	-2997.88 (8340.48)
	$N_r \uparrow$	PNAUTLR	$+16.40$ (39.71)
accretion of cloud water by rain	$Q_c \downarrow$	PQACCL	-2.27×10^{-8} (3.64×10^{-8})
	$Q_r \uparrow$	PQACCL	$+2.27 \times 10^{-8}$ (3.64×10^{-8})
	$N_c \downarrow$	PNACCL	-22002.6 (29000.3)
self-collection of raindrop	$N_r \downarrow$	PNSCLL	4.220 (4.823)
aggregation of ice	$Q_i \downarrow$	PQAUTI	-1.60×10^{-8} (2.96×10^{-8})
	$Q_s \uparrow$	PQAUTI	$+1.60 \times 10^{-10}$ (2.96×10^{-8})
	$N_i \downarrow$	PNAUTI	-1.70 (3.12)
accretion of cloud by snow	$Q_c \downarrow$	PQACCS	-2.88×10^{-9} (5.70×10^{-9})
	$Q_s \uparrow$	PQACCS	$+2.88 \times 10^{-9}$ (5.70×10^{-9})
	$N_c \downarrow$	PNACCS	-3344.7 (5718.9)
accretion of ice by snow	$Q_i \downarrow$	PQACCIS	-5.23×10^{-11} (1.41×10^{-10})
	$Q_s \uparrow$	PQACCIS	$+5.23 \times 10^{-11}$ (1.41×10^{-10})
	$N_i \downarrow$	PNACCIS	-3.83 (9.98)
accretion of rain by snow	$Q_r \downarrow$	PQACCRS	-1.79×10^{-11} (3.49×10^{-11})
	$Q_s \uparrow$	PQACCRS	$+1.79 \times 10^{-11}$ (3.49×10^{-11})
	$N_r \downarrow$	PNACCRS	-1.88×10^{-5} (2.33×10^{-5})
Hallet-Mossop process	$Q_c \downarrow$	PQHLMi	-6.66×10^{-13} (2.07×10^{-12})
	$Q_i \uparrow$	PQHLMi	$+6.66 \times 10^{-13}$ (2.07×10^{-12})
	$N_i \uparrow$	PNHLMi	0.32 (0.99)
self-aggregation of snow	$N_s \downarrow$	PNSAGS	-0.12 (0.28)
Bergeron process (ice)	$Q_c \downarrow$	PBERGI	-7.28×10^{-9} (8.46×10^{-9})
	$Q_i \uparrow$	PBERGI	$+7.28 \times 10^{-9}$ (8.46×10^{-9})
Bergeron process (snow)	$Q_c \downarrow$	PBERGS	-1.67×10^{-8} (2.24×10^{-8})
	$Q_s \uparrow$	PBERGS	$+1.67 \times 10^{-8}$ (2.24×10^{-8})

project (Dal Gesso et al., 2015), that contributes to understand the low cloud feedback (e.g., Zhang et al., 2012b, 2013). The experiments are based on the meteorological condition of the CGILS S11 warm rain case (stratocumulus; Sc). The simulations are run for 30 days and the last 20 days were analyzed. The further descriptions of the CGILS project are provided elsewhere (e.g., Blossey et al., 2013; Bretherton et al., 2013; Dal Gesso et al., 2015).

Figure 5.4 shows the vertical profiles of cloud water and rainwater contents simulated in the SCM. The results from DSD experiments (Figs. 5.4a and 5.4b) illustrate wide spread in the profile of rainwater depending on the shape parameter μ whereas that of cloud water is quite similar in any prescribed DSD. The modified gamma distribution with $\mu = 10$ shows the smallest rainwater, whereas the exponential DSD (i.e., Marshal-Palmer distribution; shown as the red curve) represents the largest value.

Next, the sensitivity of hydrometeor profiles on the autoconversion schemes are examined (Figs. 5.4c and 5.4d) with shape parameter $\mu = 5$ since the exponential distribution ($\mu = 1$)

Table 5.4: Sensitivity experiments with respect to autoconversion scheme, raindrop size distribution (DSD), and sub-time step applied in the prognostic precipitation scheme.

Experiment Name/Type	Autoconversion Scheme	Rain Size Distribution	Sub-time step
DSD Experiment			
1) PROG 2M	Berry (1968)	Marshall-Palmer	120 sec
2) PROG 2M- $\mu 2$	Berry (1968)	Modified-Gamma ($\mu = 2$)	120 sec
3) PROG 2M- $\mu 5$	Berry (1968)	Modified-Gamma ($\mu = 5$)	120 sec
4) PROG 2M- $\mu 10$	Berry (1968)	Modified-Gamma ($\mu = 10$)	120 sec
Microphysics Experiment			
5) PROG 2M-KK00	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	120 sec
6) PROG 2M-Be94	Beheng (1994)	Modified-Gamma ($\mu = 5$)	120 sec
7) PROG 2M-TC80	Tripoli and Cotton (1980)	Modified-Gamma ($\mu = 5$)	120 sec
8) PROG 2M-LD04	Liu and Daum (2004)	Modified-Gamma ($\mu = 5$)	120 sec
Iteration Experiment			
9) PROG 2M-1s	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	1 sec
10) PROG 2M-30s	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	30 sec
11) PROG 2M-60s	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	60 sec
12) PROG 2M-300s	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	300 sec
13) PROG 2M-600s	Khairoutdinov and Kogan (2000)	Modified-Gamma ($\mu = 5$)	600 sec

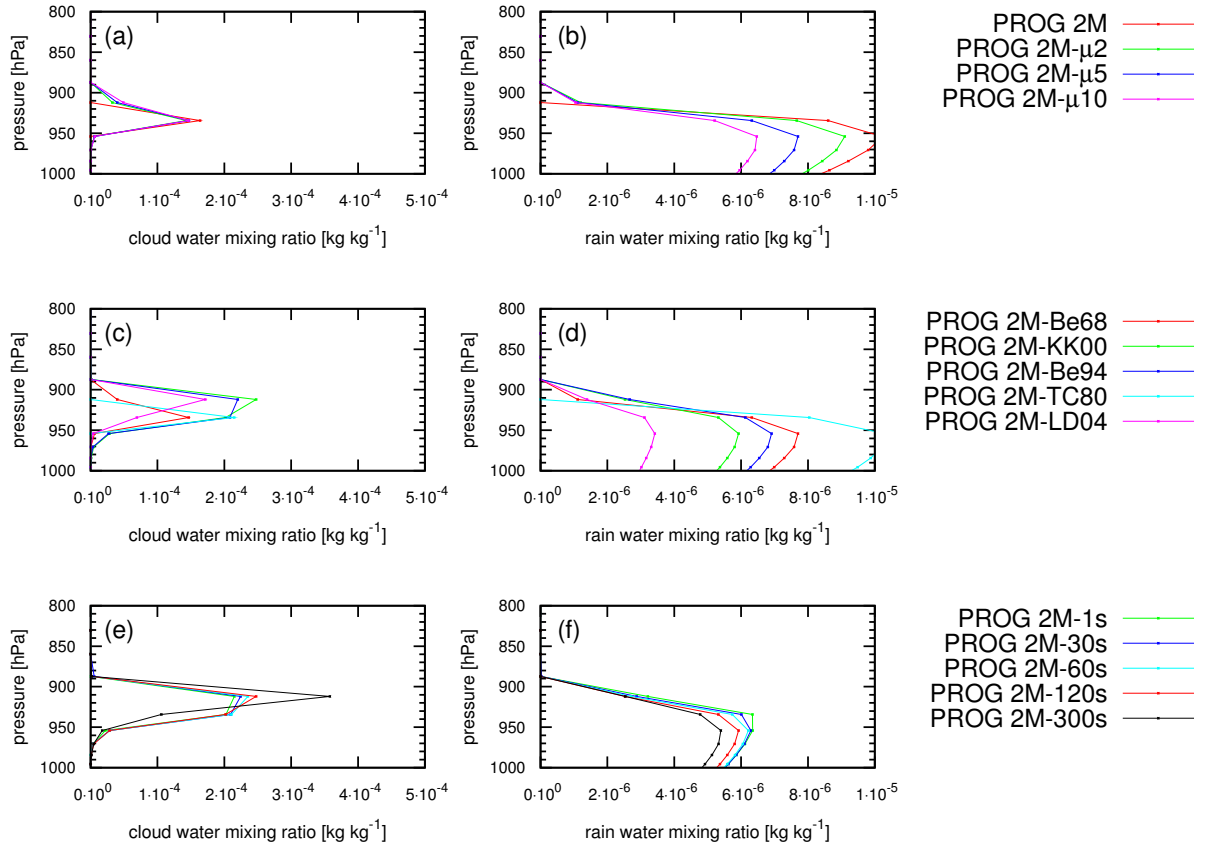


Fig. 5.4: The vertical profiles of cloud water and rainwater mixing ratios based on the sensitivity experiments for (a, b) DSD, (c, d) autoconversion schemes, and (e, f) sub-time step using SCM in CGILS S11 case.

overestimates rainwater and $\mu = 10$ is unrealistic (Posselt and Lohmann, 2009; Abel and Boutle, 2012). The large spread among the scheme is observed because autoconversion controls the ratio of cloud water to rainwater, thus cloud lifetime (e.g., Suzuki et al., 2013a; Gettelman et al., 2013). As discussed in Michibata and Takemura (2015), the scheme proposed by Tripoli and Cotton (1980, hereinafter TC80) diagnoses high amount of rainwater whereas those by Liu and Daum (2004, hereinafter LD04) and Khairoutdinov and Kogan (2000, hereinafter KK00) show relatively smaller rainwater. The scheme proposed by Beheng (1994, hereinafter Be94) diagnoses moderate values among the five schemes that is sandwiched by KK00 and Berry (1968, hereinafter Be68). Be94 scheme depends strongly on N_c (i.e., the value of β in Eq. 5.1), which will cause too strong aerosol forcing due to the highest nonlinear dependence (e.g. Michibata and Takemura, 2015; Suzuki et al., 2013a). Be68 autoconversion scheme is the default in MIROC DIAG scheme, and the sensitivity of the different use of autoconversion scheme to global simulations will be discussed later continuously (cf. Section 5.4), using Be68 and KK00 schemes.

Finally, the sensitivity in the number of sub-time steps is evaluated (Figs. 5.4e and 5.4f). Note that the length of sub-time step is not a tuning parameter because the “exact” solution of the prognostic equations cannot be obtained from finite time steps. The so-called “numerical error” is unavoidable in numerical simulations. Given that the simulation using sub-time step with 1 sec is a solution with the smallest error in the experiments here, simulations whose sub-time step are 300 sec (black solid) and 120 sec (red solid) have visible biases relative to the simulations whose sub-time step smaller than 60 sec.

In summary above, the KK00 autoconversion scheme with modified gamma DSD ($\mu = 5$) using 60 sec sub-time step for cloud microphysics iteration is the standard configuration for a global simulation discussed below.

5.4 Global Model Simulations

To understand the impact of the PROG scheme on climate simulation in GCM, we perform global simulations using MIROC-SPRINTARS (Watanabe et al., 2010) with both DIAG and PROG configurations. The horizontal and vertical model resolutions are $2.8^\circ \times 2.8^\circ$ in longitude and latitude, and 20 vertical sigma layers, respectively (T42L20). The standard model time step is 12 min for both DIAG and PROG experiments. Simulations are conducted in the present-day condition using monthly mean climatological sea surface temperature (SST). Experiments are integrated for 6 years, and are analyzed the last 5 years. Note that the solid precipitation is still

a diagnostic treatment here, since the aims of this Chapter is to evaluate how the new scheme improve the warm rain microphysics and thus cloud-radiation-climate feedbacks. Precipitating hydrometeors (i.e., rainwater and snowflake) are not considered in a radiation calculation for both DIAG and PROG configurations.

5.4.1 Rain Formation Process

Figure 5.5 shows global distribution of process rates with regard to autoconversion and accretion from DIAG and PROG schemes. As we expected, the contribution of the accretion process significantly increase in the PROG scheme since the PROG can memory the history of rainwater in the atmosphere across multiple time steps. The relative contribution of accretion is approximately half an order of magnitude lower in DIAG simulations, whereas that is about an order of magnitude higher in PROG (see also Table 5.5). This change of the balance of autoconversion and accretion in PROG is reasonable for other prognostic precipitation GCMs (CAM5 by Gettelman et al. (2015) and ECHAM5 by Sant et al. (2015)), which is in good agreement with the VOCALS in-situ aircraft measurement (Wood et al., 2011; Gettelman et al., 2013). The vertical distributions of process rates also change significantly (Fig. 5.6). The height where new raindrop is generated by autoconversion is lower in the PROG scheme relative to DIAG simulation (Figs. 5.6a and 5.6b). The accretion rate increases in the whole layers in the PROG, and especially higher values are found near the surface. This is not clear in the DIAG scheme mainly due to the omission of sedimentation process without the memory of rainwater.

As a result of the collection process above and sedimentation, the rainwater mass and number concentrations are higher near the ground (Fig. 5.7). In particular, the raindrop number concentration (N_r) varies very widely, from $\sim 10^2$ to $\sim 10^5$ m^{-3} in zonal annual mean, and high concentration of N_r especially appears below 800 or 900 hPa. The numerous raindrops are observed in the typical Sc deck, off the west coasts of California and Peru for example (not shown), where drizzling shallow precipitation is frequently observed (Wood, 2012). Sc cloud system is known to form drizzle frequently whose occurrences are ~ 20 – 50 % (Leon et al., 2008; Michibata et al., 2014; Wu et al., 2015). Drizzle particles have important roles in the enhancement of collision-coalescence for cloud droplets with drizzle and raindrops, thus cloud lifecycle (Wood, 2005a) as well as shortwave radiation (Muhlbauer et al., 2014).

It should be emphasized that the model with PROG scheme can reproduce the geographical distribution of drizzle very well as shown in Fig. 5.8a. Note here that the PROG scheme assumes

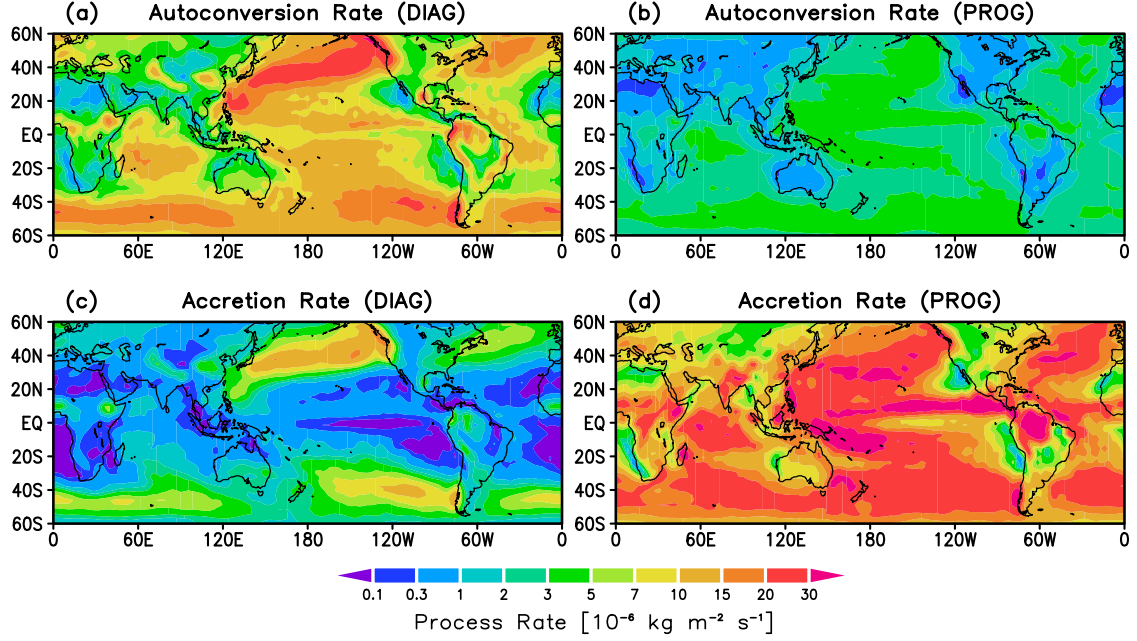


Fig. 5.5: Geographical distributions of annual mean (a, b) vertically-integrated autoconversion rate and (c, d) vertically-integrated accretion rate (unit: $10^{-6} \text{ kg m}^{-2} \text{ s}^{-1}$) for DIAG and PROG schemes.

the new generated raindrop has its size of $25 \mu\text{m}$ in radius following Khairoutdinov and Kogan (2000), which means the model treats drizzle as the rain class. Small drizzle droplets whose effective radii are less than $50 \mu\text{m}$ appear off the Californian, Peruvian, Australian, Namibian, and Canarian. The simulated rain water path (RWP) is also consistent with CloudSat observation (Li et al., 2008, see also errata, Li et al. (2011)) and multiscale model simulation (Chern et al., 2016), as shown in Fig. 5.8b and Table 5.5. Since the droplet size is determined by both rain mass and number mixing ratios, two-moment approach in the PROG scheme is working to represent hydrological system in good agreement with observations (e.g., Wood, 2005b; Wang et al., 2009; Wu et al., 2015). This will also provide advantages in representation of radiative properties as discussed in the following section.

5.4.2 Cloud-Radiation-Climate Interactions

Introduction of the prognostic equations for rainwater into the model significantly alters the distributions of liquid cloud water and cloud ice as shown in Figs. 5.9 and 5.10. More specifically, additional prognostic hydrometeor of rain reduces cloud water because rainwater converted from cloud water is reserved in the atmosphere across time steps. Ice hydrometer is also reduced because a source of the ice class from cloud water decreases due to the shift to the rain class. A

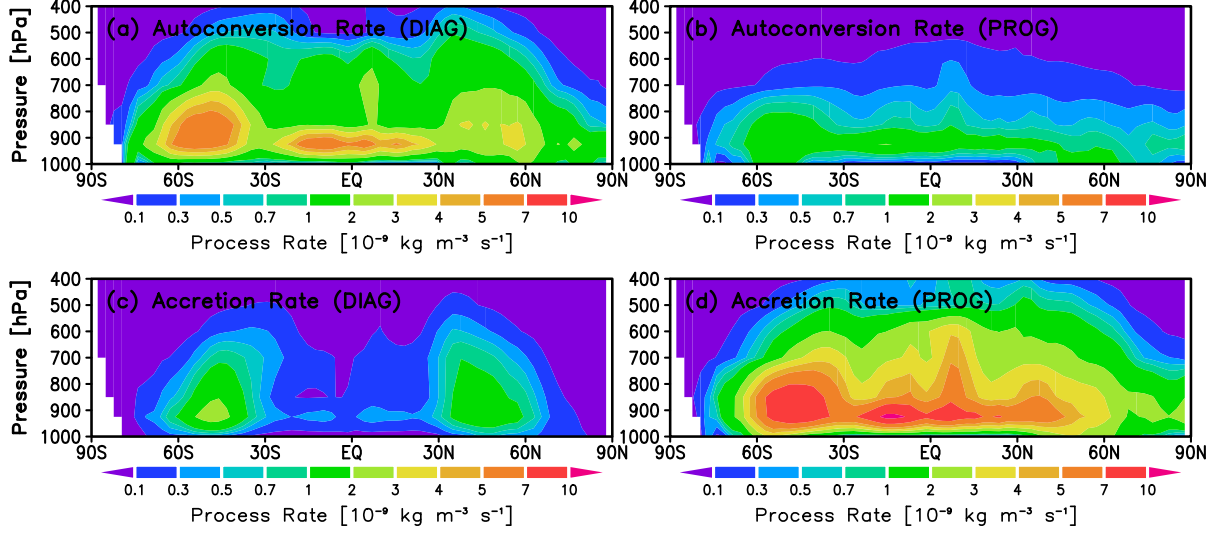


Fig. 5.6: Annual zonal mean distributions of (a, b) autoconversion rate and (c, d) accretion rate (unit: $10^{-9} \text{ kg m}^{-3} \text{ s}^{-1}$) for DIAG and PROG schemes.

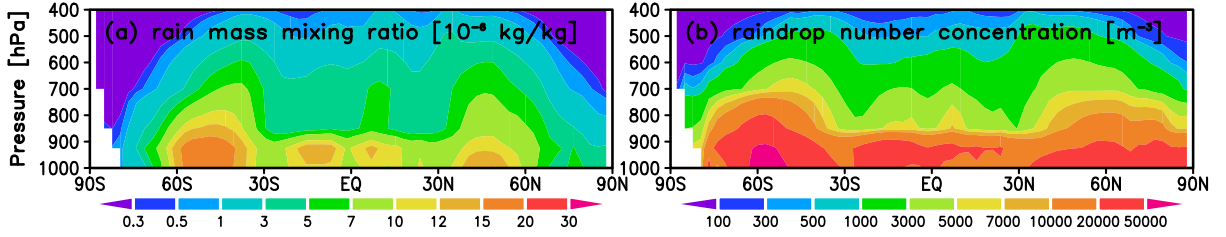


Fig. 5.7: Annual zonal mean distributions of mixing ratios for (a) rainwater mass ($10^{-6} \text{ kg kg}^{-1}$) and (b) raindrop number concentration (m^{-3}) simulated by the PROG scheme.

hydrometeor class of snow is therefore one of the important future work to avoid the inconsistency between liquid and solid precipitation by online coupling among five hydrometeor classes (i.e., water vapor, cloud water, cloud ice, rainwater, and snow), namely, prognostic snow framework.

Figures 5.9a and 5.9b illustrate systematic reduction of cloud water in the PROG simulation at height of 500–800 hPa especially, and cloud water distributes lower altitude relative to the DIAG simulation. This would be caused by not only water conversion from cloud to rain, but also an evaporation of rainwater during sedimentation (i.e., drizzle evaporation), which alter the vertical distribution of water vapor as well as cloud water for lower altitude. Such the systematic shift of cloud water also changes the vertical distribution of ice hydrometeor (Figs. 5.9c and 5.9d), because the source term of ice from liquid water in higher altitude is depleted. This is also evident in horizontal distribution of column water path shown in Fig. 5.10. The global annual mean LWP is decreased by switching from the DIAG to PROG scheme and is consistent with MODIS and

Table 5.5: Simulated global mean climatology of physical parameters for DIAG and PROG schemes. Observational data are obtained from GPCP version 2.2 (for the period 2001–2013) for precipitation rate, ISCCP D2 (for the period 2005–2009) for cloud cover, Aqua-MODIS Level 3 (for the period 2006–2010) for cloud optical thickness (COT), cloud-top effective radius (CER), and liquid water path (LWP).

	DIAG		PROG		Obs.
	Be68	KK00	Be68	KK00	
Autoconversion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	8.5×10^{-6}	7.8×10^{-6}	2.1×10^{-6}	2.2×10^{-6}	–
Accretion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	1.9×10^{-6}	2.1×10^{-6}	1.6×10^{-5}	1.6×10^{-5}	–
Total conversion rate ($\text{kg m}^{-2} \text{s}^{-1}$)	1.0×10^{-5}	9.9×10^{-6}	1.8×10^{-5}	1.9×10^{-5}	–
Precipitation rate (mm day^{-1})	3.04	3.03	2.25	2.23	2.67
Precipitation by Cumulus (mm day^{-1})	1.57	1.59	1.03	1.02	–
Precipitation by LSC (mm day^{-1})	1.47	1.44	1.22	1.21	–
Cloud cover (%)	60.25	62.58	51.32	51.74	65.4 ^a
COT	10.4	12.5	7.9	7.9	13.1 ^a
CER liquid (μm)	11.7	11.8	11.3	11.2	13.8 ^a
CER ice (μm)	81.2	80.2	197.1	196.9	24.2 ^b
CLWP (g m^{-2})	87.5	107.4	54.7	54.7	84.7 ^c
					31.6–58 ^d
CIWP (g m^{-2})	23.9	24.1	10.8	10.8	23.0–26.7 ^e
RWP (g m^{-2})	12.2	12.0	28.9	30.2	26.7 ^f

^aAveraged from 60°S to 60°N

^bEffective radius of ice crystal taken from MODIS used in Table 4 of Barahona et al. (2014).

^cSSM/I data (for the years 2006–2010) averaged from 60°S to 60°N only for ocean.

^dMODIS and AMSR-E retrievals by Lebsock and Su (2014).

^eCloudSat 2B-CWC, 2C-ICE, and DARDAR products as given in Li et al. (2012).

^fCloudSat FLAG 2B-CWC Product as given by Li et al. (2008).

AMSR-E observations (Lebsock and Su, 2014), while the ice water path (IWP) is significantly underestimated in PROG compared to the observations (Li et al., 2012; Waliser et al., 2009) as shown in Table 5.5. This will be addressed in future study by adding prognostic snow.

Figure 5.11 shows global distribution of surface precipitation rate from DIAG and PROG schemes and observations. The observational data is taken from Global Precipitation Climatology Project (GPCP) observations version 2.2 (Adler et al., 2003) for the period 2001–2013. Although the geographical distribution is quite similar between MIROC simulations and GPCP observations, the traditional DIAG scheme overestimates the precipitation by about 13.5 %. This could be related to the issue “too fast, too frequent rain formation” (Suzuki et al., 2013a) which is a common issue in GCMs (Suzuki et al., 2015). As illustrated in Fig. 5.10, the standard DIAG has too large amount of cloud water, which causes high autoconversion rate and thus too large precipitation. On the other hand, the PROG simulations underestimate the global mean precipitation rate by between 15.7–16.5 %, while the overestimates over Indian Ocean and east of the Philippines are significantly improved. Furthermore, relatively weak rainfall regions, off the west coast of California, Peruvian, and Canarian are also well captured in PROG simulation.

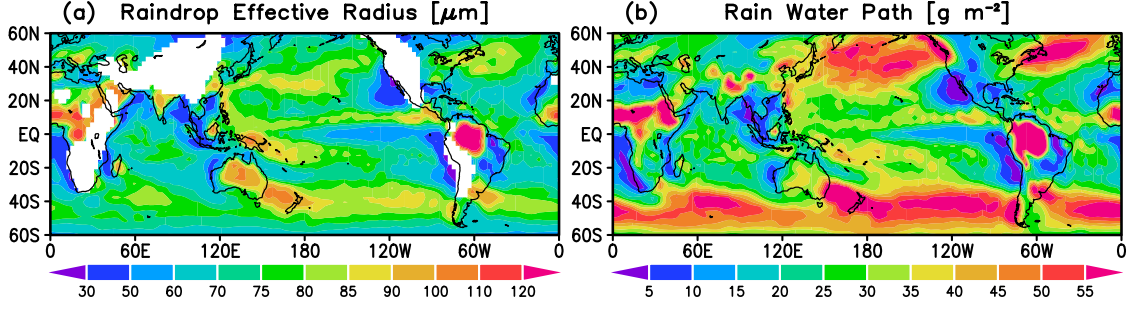


Fig. 5.8: Geographical distributions of annual mean (a) raindrop effective radius (μm) at 900 hPa and (b) rain water path (g m^{-2}) based on simulations with PROG scheme.

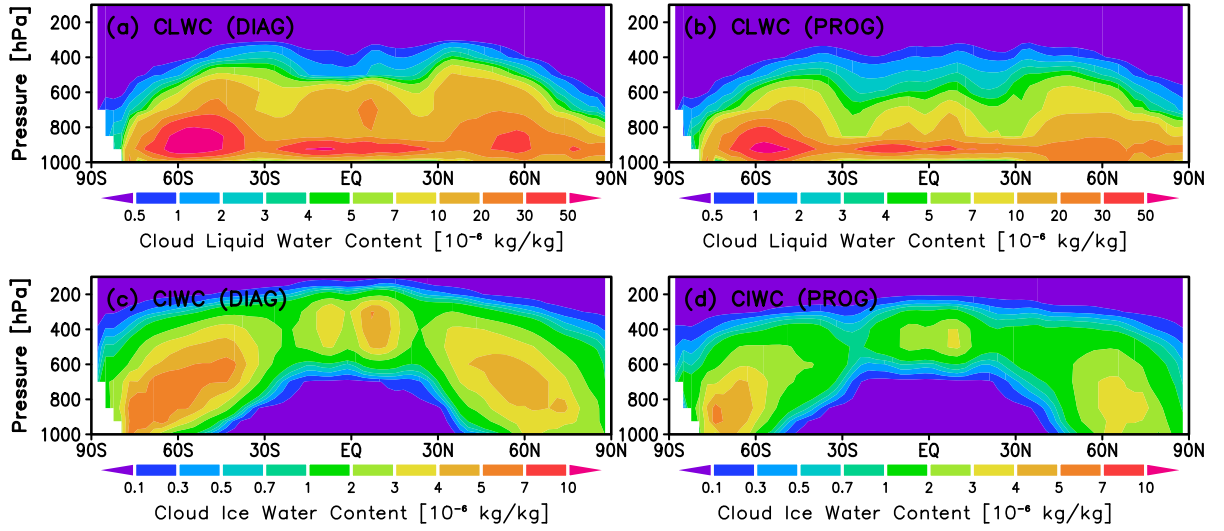


Fig. 5.9: Annual zonal mean distributions of mass mixing ratios for (a, b) cloud liquid water and (c, d) cloud ice water (unit: $10^{-6} \text{ kg kg}^{-1}$) for DIAG and PROG schemes.

The PROG scheme can consider the interaction with cloud and rain (accretion), because it stores rainwater in the atmosphere, which would improve the timescale of hydrological cycle. Although the MODIS/CloudSat satellite simulators (Bodas-Salcedo et al., 2011) are powerful tools to examine the cloud-to-rain transition at a fundamental process-level, it is not available here due to their high computational costs. The both DIAG and PROG simulations do not represent ITCZ strong precipitation belt, which implies not only microphysics but also other processes of the model (e.g., boundary processes and physics within subgrid scale) still have more fundamental errors (Michibata and Takemura, 2015).

Finally, we examine the changes in radiative properties depending on the treatment of rain. Figure 5.12 shows global distribution of shortwave cloud radiative forcing (SWCRF) for model and satellite estimates. The observational data is taken from the Clouds and Earth's Radiant

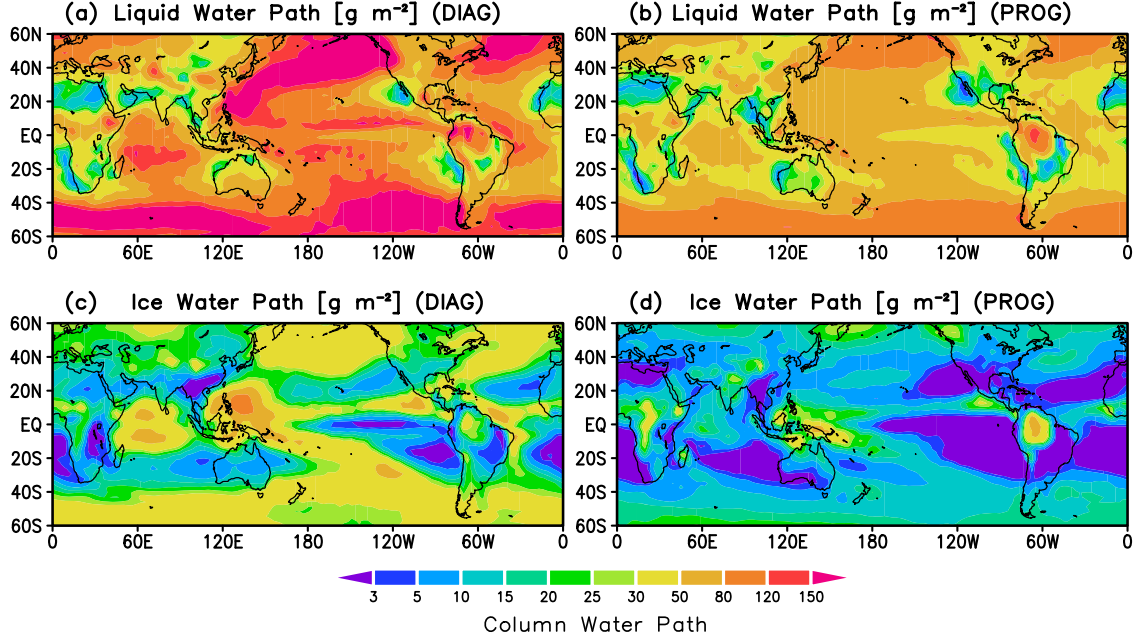


Fig. 5.10: Geographical distributions of annual mean (a, b) cloud liquid water path and (c, d) cloud ice water path (unit: g m^{-2}) for DIAG and PROG schemes.

Energy Systems (CERES) Energy Balanced and Filled (EBAF) TOA Edition 2.8 (Loeb et al., 2009) for the period 2001–2013. Simulation with the DIAG scheme substantially underestimates the SWCRF over the Indian Ocean, western pacific, and ITCZ/SPCZ. This “too bright” low-cloud bias causes overestimation of -5 to -10 W m^{-2} from CERES observation (-47.2 W m^{-2}) in global mean. It should be noted that the PROG simulation improves the bias significantly, and its global mean (-44.2 W m^{-2}) is very close to the observation. In the PROG scheme, cloud droplets and rain drops coexist in the cloud, which is more realistic representation. However, it is not true in the traditional DIAG simulation owing to the ad hoc treatment of rain. Although the longwave CRF is not evaluated here because this Chapter aims to improve precipitation in liquid phase, it will be discussed when the prognostic snow scheme is introduced additionally.

5.5 Summary Perspectives

In this Chapter, a physically based two-moment bulk microphysics scheme was developed. Prognostic equations for rainwater mass and number mixing ratios are newly inserted into an LSC parameterization, and now the rainwater can be stored in the atmosphere across multiple time steps. The performance of the new scheme (PROG) was evaluated in a box model, SCM, and global model.

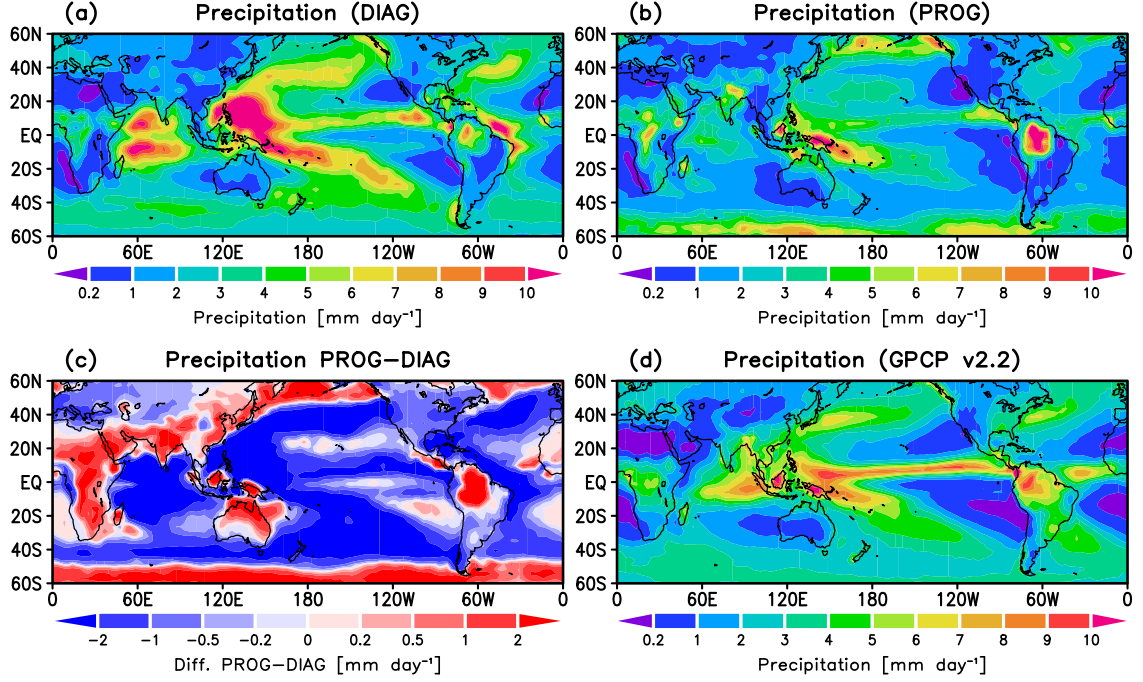


Fig. 5.11: Geographical distributions of annual mean precipitation rate (mm day^{-1}) for (a) DIAG, (b) PROG, (c) the difference between PROG and DIAG, and (d) GPCP version 2.2 observation (averaged over the years 2001–2013).

Firstly we examined the mass and number budget among the four prognostic hydrometeor classes (i.e., liquid cloud water, cloud ice, rainwater, and snow) using box model under idealized conditions. The prognostic treatment of rain significantly increased the relative contribution of accretion process against to that of autoconversion, which is a more realistic signal of cloud-precipitation microphysics. The SCM test implied that the set of best combination of KK00 autoconversion scheme with modified gamma DSD ($\mu = 5$) using 60 sec sub-time step iteration for microphysics.

Then global simulations were conducted using an aerosol-climate model, MIROC-SPRINTARS GCM, with the configurations recommended from the above simple model experiments. Simulation with the PROG scheme represented the horizontal distribution of coastal stratocumuli very well. The model simulated their effective radius with 30–60 μm , which is consistent with in-situ and satellite observations for drizzling stratocumulus (e.g., Wood, 2005b; Leon et al., 2008; Wang et al., 2009; Wu et al., 2015). Inserting the additional hydrometeor class of rainwater decreased liquid cloud water, whose value in global annual mean was close to the MODIS and AMSR-E satellite observations (Lebsock and Su, 2014). This further improved the horizontal distributions of precipitation rate and SWCRF. The traditional DIAG scheme underestimates the SWCRF

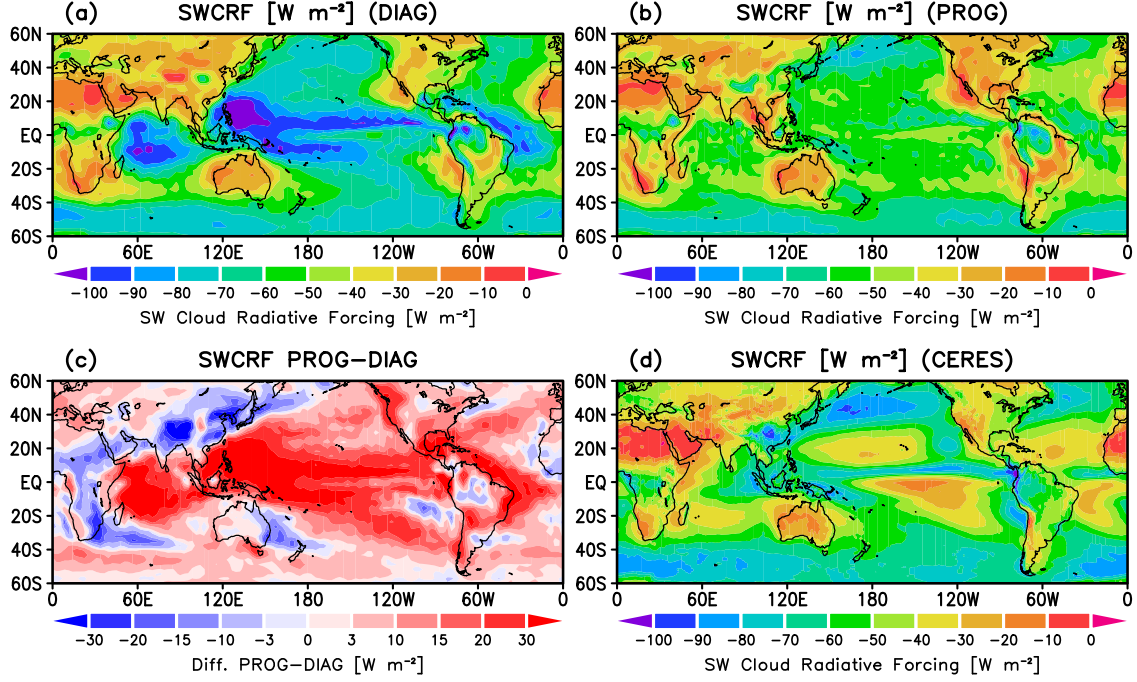


Fig. 5.12: Geographical distributions of annual mean shortwave cloud radiative forcing (SWCRF; W m^{-2}) for (a) DIAG, (b) PROG, (c) the difference between PROG and DIAG, and (d) CERES EBAF-TOA Ed2.8 observation (averaged over the years 2001–2013).

(i.e., “too bright” property for low-clouds) approximately -5 to -10 W m^{-2} in global mean, and the bias was pronounced over the Indian Ocean and western Pacific. The PROG simulation significantly reduced the bias in SWCRF, and was very close to the CERES satellite estimate. These improvements in the PROG simulation could be a result from more physically based representation of cloud and precipitation, namely, the cloud droplets and raindrops coexist in the cloud, while this is not true in the traditional diagnostic treatment of rain, which is a bottleneck in GCMs (e.g., Wang et al., 2012; Gettelman et al., 2013; Michibata and Takemura, 2015; Suzuki et al., 2015).

However, some deteriorations were also found in the new scheme. For example, depletion of cloud liquid water by rainwater resulted in the vertical distribution of water mass for lower altitude, which caused the reduction of source term of cloud ice. As a result, the PROG scheme had biases in the vertical distribution of ice cloud and the IWP larger than the default configuration of the MIROC. This will be improved in future study by adding prognostic snow scheme, which is more sophisticated cloud-precipitation framework.

Overall, this Chapter achieved the development of advanced cloud-precipitation microphysics scheme, and it emphasized the importance of the prognostic treatment of rain in GCMs. These

results suggest that the prognostic rain framework will promise for better process representations of aerosol-cloud-precipitation-radiation interactions, which will overcome the common problems of GCMs such the “too few too bright low-cloud problem” (Nam et al., 2012) and “too rapid too frequent rain formation” (Suzuki et al., 2015). Improvements in the ACI using the new PROG scheme will be further discussed in the next Chapter.

6 Improved Representation of Aerosol-Cloud-Precipitation Interactions with a Use of Prognostic Precipitation Scheme

Abstract

General circulation models (GCMs) still have common problems in their representation of cloud-precipitation microphysics and the magnitude of aerosol-cloud interactions (ACI). One of the main causes is their diagnostic treatment of rain. Here we investigate how treatment of precipitation in GCMs improve the regime dependence of ACI for warm clouds, and examine the mechanisms, by employing the both prognostic and diagnostic treatment of rain in MIROC-SPRINTARS aerosol-climate model. We find that both increasing and decreasing responses of liquid water path (LWP) to aerosol perturbations in the prognostic rain scheme as observed in satellite observations and large-eddy simulation in recent literature, while the traditional diagnostic scheme shows only increasing LWP response so-called cloud lifetime effect. This “bidirectional” LWP responses appeared in the new scheme depend strongly on the cloud life stage which is characterized by the cloud geometric thickness and the ratio of rainwater to total liquid water in the cloud. As a result, the simulation based on the prognostic treatment of rain reduces the LWP susceptibility with a magnitude of -30% relative to the diagnostic scheme. These improvements support that the prognostic precipitation framework has great advantages in process representations of not only aerosol-cloud-precipitation microphysics but also cloud system responses to aerosol perturbations with buffering systems.

Keywords: GCM, prognostic rain scheme, aerosol-cloud-precipitation interactions

Key Points:

- This work exhibits the prognostic treatment of rain (PROG) significantly improves the cloud response to aerosol perturbations, whereas the traditional GCM with diagnostic treatment of rain (DIAG) overestimates the ACI.
- The PROG scheme captures the both increase and decrease responses of cloud water to increased aerosols, which is consistent with observations, while the DIAG scheme cannot.

- To implement PROG scheme into GCMs has great advantages for more accurate climate simulation with the realistic representation of the ACI.
-

6.1 Introduction

Atmospheric aerosols play significant roles in the climate system via their interaction with clouds (Twomey, 1977; Albrecht, 1989), so-called aerosol-cloud interactions (ACI). Since warm clouds in the lower troposphere are particularly susceptible to aerosols (e.g., Pincus and Baker, 1994; Bréon et al., 2002; Kubar et al., 2009) as well as surrounding environmental conditions (e.g., Sorooshian et al., 2013; Chen et al., 2014), modeling the aerosol-cloud-precipitation-climate feedback in general circulation models (GCMs) is one of the most uncertain processes and remains controversial (Seinfeld et al., 2016; Fan et al., 2016).

Such the marine stratiform warm clouds contribute to cooling the climate system. They frequently form drizzle with the occurrences in $\sim 20\text{--}50\%$ (Leon et al., 2008; Kubar et al., 2009; Michibata et al., 2014; Wu et al., 2015), which is important in the onset of precipitation via the collision-coalescence of cloud droplets with raindrops, thus cloud lifecycle (Wood, 2005a) and shortwave radiation (Muhlbauer et al., 2014). The contribution of the aerosol-cloud-precipitation interaction in the Earth’s radiation budget is therefore significant, even though the changes in cloud fraction is small (Wood, 2012).

Almost all of GCMs, however, treat precipitation diagnostically, which means the produced rainwater falls out to the surface instantaneously. Although this is a suitable handling in GCMs with long time step assuming large fall velocity of precipitating hydrometeors, the diagnostic treatment of rain (hereafter “DIAG”) misses smaller raindrops such a drizzle which does not reach to the surface within one time step. This will further cause biases in precipitation frequency and intensity (Stephens et al., 2010; Suzuki et al., 2013a, 2015) as well as the cloud brightness and lifetime (Nam et al., 2012; Quaas et al., 2009; Wang et al., 2012). Suzuki et al. (2013b) revealed that the inconsistency between “process-based constraints” of microphysics estimated from satellite and “performance-oriented constraints” of the climate simulation with empirical model tuning. This suggests that GCMs evade the biases in the modeled microphysical parameterization and climate trends, due to their error compensations. Michibata and Takemura (2015) performed sensitivity experiments using five autoconversion schemes (mass transfer from cloud water to rainwater) in MIROC-SPRINTARS, and reported that the significant improvements with well-

balanced both cloud microphysics and radiative property were not found in any schemes. These studies strongly claim the necessity of improvements in the aerosol-cloud-precipitation framework with more realistic physical parameterization at a fundamental process-level (Seinfeld et al., 2016; Michibata et al., 2016).

In recently, advanced cloud-precipitation microphysics framework was developed (Gettelman and Morrison, 2015; Sant et al., 2015) with a prognostic treatment of rain (hereafter “PROG”). They succeeded to represent more realistic cloud-precipitation microphysics and also improved their radiative forcing. More specifically, the balance of autoconversion (collisional growth among cloud droplets) and accretion (collision-coalescence of cloud droplets by larger raindrops) was significantly improved due to the enhancement of accretion process, which is consistent with observations. Since the aerosol effects depend on the autoconversion process via cloud droplets number concentration N_c , rather than on the accretion process which is parameterized by only masses of cloud water and rainwater, enhancement of the accretion over the autoconversion contributes to reduce the aerosol forcing (Posselt and Lohmann, 2009; Gettelman et al., 2013).

There is an additional problem in GCMs. That is, GCMs with long time step and coarse resolution cannot capture the feedback of cloud microphysics onto environment and dynamics (Michibata et al., 2016; Zhou and Penner, 2017), which plays important roles within a shorter timescale than GCMs resolving. To implement the so-called “microphysics-dynamics interactions (Wood, 2012; Michibata et al., 2016)” which buffer the aerosol perturbations via entrainment mixing (Small et al., 2009) and/or precipitation enhancement (Rosenfeld et al., 2014) into GCMs is still one of the most challenging issues (e.g., Stevens and Feingold, 2009; Tao et al., 2012; Seinfeld et al., 2016).

This may lead uncertainty in responses of cloud liquid water path (LWP) either increase or decrease against the increased aerosols, thus related to the cloud lifetime effect (Guo et al., 2015). Michibata et al. (2016) showed that the response of LWP to aerosol perturbations is strongly related to cloud types and thermodynamic conditions, and hence the clear regional dependence was found from the A-Train satellite data for both positive and negative LWP susceptibilities. The positive LWP response was found over midlatitude where stratiform clouds are dominant, due to effective condensation with sufficient water vapor in the stable environment. On the other hand, the negative response was appeared over tropics/subtropics where more cumulative clouds are dominant, due to entrainment of dry air and evaporation in the unstable environment. This “regime dependence” microphysics-dynamics interaction could not be captured by MIROC-

SPRINTARS. We would like to emphasize that the fundamental flaw in the LWP susceptibility will be applied not only in our model but also in other GCMs as a common problem because the modeled autoconversion scheme increases cloud water with increasing aerosols by assuming only the cloud lifetime effect without buffering processes.

An implementation of the PROG scheme into GCM will be a more realistic physics-based framework for aerosol-cloud-precipitation-climate interactions, while its impacts on cloud responses to aerosol perturbations is poorly understood. To this end, we investigate the impact of PROG scheme on cloud-precipitation microphysics with a main focus on the LWP susceptibility based on different cloud regimes, and thus, aim to suggest the advantages in process representation of the PROG framework against to the traditional DIAG scheme for future model development. Note that the present study focuses only on warm-phase low clouds (cf. Model and Methods).

6.2 Model and Methods

6.2.1 General Model Description and the Prognostic Rain Scheme

We employed global aerosol-climate model, MIROC-SPRINTARS (Watanabe et al., 2010), the latest version 6. The model explicitly predicts the mass mixing ratios of main aerosol species in the troposphere (i.e., black carbon, organic matter, sulfate, soil dust, and sea salt) and precursor gases of sulfate (Takemura et al., 2000, 2005).

The cloud microphysics scheme is PDF-based prognostic large-scale condensation (LSC) scheme (Watanabe et al., 2009), which considers subgrid-scale variability of temperature and total water content. The both cloud water and cloud ice are treated as two-moment representation by prognosing mass and number mixing ratios (Takemura et al., 2009). Since the standard version of MIROC-SPRINTARS does not prognose the rainwater, the diagnosed precipitation due to autoconversion is removed instantaneously from the atmosphere as surface precipitation within a single time step (Ghan and Easter, 1992).

We have newly introduced mixing ratios of rainwater q_r [kg kg^{-1}] and raindrop number concentration N_r [m^{-3}] as prognostic variables based on a two-moment bulk microphysics into the LSC scheme. The both rain mass and number mixing ratios are stored in the atmosphere across multiple time steps, and they vary with processes due to evaporation, autoconversion, accretion, self-collection, and sedimentation. The Khairoutdinov and Kogan (2000) parameterization for autoconversion and accretion, and Beheng (1994) self-collection scheme are implemented as a

standard configuration of the new prognostic rain scheme into MIROC. The subgrid fraction of precipitation is also considered by assuming the precipitating hydrometeor falls from overlapping clouds in vertical, so-called maximum overlap.

The above microphysical processes (i.e., evaporation, autoconversion, accretion, self-collection, and sedimentation) are calculated within the sub-time step (60 sec in this study), and the iteration with regard to the sedimentation of rainwater can be shorter subject to the vertical CFL condition. The raindrop size distribution $N(D)$ is based on the modified-gamma distribution (Ulbrich, 1983), which is described as,

$$N(D) = \frac{N_r}{\Gamma(\mu)} \left(\frac{D}{D_0} \right)^\mu \frac{1}{D} \exp \left(-\frac{D}{D_0} \right) \quad (6.1)$$

where $\Gamma(\mu)$ is the Gamma function. N_r and D are the total raindrop number concentration [m^{-3}] and the raindrop diameter [m], respectively. $N(D)$ is controlled by both the shape parameter μ and the distribution parameter D_0 described as,

$$D_0 = \frac{\bar{D}}{\sqrt[3]{(\mu+2)(\mu+1)\mu}} \quad (6.2)$$

where the mean diameter of raindrop $\bar{D}$ [m] is,

$$\bar{D} = \sqrt[3]{\frac{6}{\pi \rho_w} \frac{\rho_a q_r}{N_r}} \quad (6.3)$$

ρ_a and ρ_w are the densities [kg m^{-3}] with regard to the air and water, respectively. We use $\mu = 5$ for raindrop size distribution following Sant et al. (2015).

The mass and number weighted bulk fall velocities v_q and v_N , which are the approximation based on Rogers et al. (1993), are described as follows (Posselt and Lohmann, 2009; Sant et al., 2015):

$$v_q = \begin{cases} (\mu+3) b_v D_0 & \text{for } D_0 < D_v/(\mu+3) \\ b_1 & \text{for } D_0 > D_v/(\mu+3) \end{cases} \quad (6.4)$$

$$v_N = \begin{cases} \mu b_v D_0 & \text{for } D_0 < D_v/\mu \\ b_1 & \text{for } D_0 > D_v/\mu \end{cases} \quad (6.5)$$

with the constants $b_1 = 9.65 [\text{m s}^{-1}]$, $b_v = 3918 [\text{s}^{-1}]$, and $D_v = 2463 [\mu\text{m}]$. Then the sedimentation of q_r and N_r in flux forms are expressed as,

$$\left. \frac{\partial q_r}{\partial t} \right|_{sed} = \frac{1}{\rho_a} \frac{\partial}{\partial z} (\rho_a q_r v_q) \quad (6.6)$$

$$\left. \frac{\partial N_r}{\partial t} \right|_{sed} = \frac{1}{\rho_a} \frac{\partial}{\partial z} (\rho_a N_r v_N) \quad (6.7)$$

which are strictly controlled by the vertical CFL criteria described earlier.

Solid precipitation is still a diagnostic variable in this study based on Wilson and Ballard (1999) microphysics scheme, and calculated within the standard model time step (i.e., 12 min) whereas liquid precipitation is treated within the sub-time step (~ 60 sec) described above. This inconsistent between liquid and solid precipitation will be improved in future publication.

6.2.2 Experiment and Analysis Sampling

To understand the advantages of new cloud-precipitation scheme with prognostic treatment of rain against to the traditional diagnostic representation, we investigated the aerosol effects on the cloud and precipitation microphysics using the both PROG and DIAG configurations. The model resolution was T42L20 (approximately $2.8^\circ \times 2.8^\circ$ in longitude and latitude with 20 vertical layers), and standard model time step was 12 min for both simulations.

Experiments were carried out with 6 years' time integration and analyzed for the last 5 years. We used 6 hourly instantaneous model output, and extracted only warm clouds (> 273.15 K) for analysis. The samples whose cloud optical thickness less than 0.2 and LWP less than $1 \text{ [g m}^{-2}\text{]}$ were also screened out from the analysis to exclude the ambiguous retrieval.

6.3 Results and Discussions

6.3.1 LWP Susceptibility in the DIAG vs. PROG Schemes

Figure 6.1 shows a geographical distribution of LWP response to aerosol perturbations simulated from the DIAG and PROG schemes. The metric LWP susceptibility defined as $d \ln \text{LWP} / d \ln N_c$ means the strength of the cloud lifetime effect. For example, the positive value could be interpreted as the increase of LWP for incremental changes in aerosols due to decrease of the auto-conversion rate to prolong the cloud lifetime (Wang et al., 2012; Zhou and Penner, 2017). It is therefore natural to accept the model result that showing positive response across the globe. We find that the $d \ln \text{LWP} / d \ln N_c$ decreases globally from 0.31 (DIAG) to 0.23 (PROG) in non-precipitating clouds, and the decrease is more obvious in precipitating clouds that from 0.35 (DIAG) to 0.22 (PROG). This means the ACI is weakened by the prognostic treatment of precipitation (approximately the range between -25.3% and -36.4%), which is linked with the occurrence of precipitation (Mülmenstädt et al., 2015) as discussed later. The signal of decreasing ACI shown in the PROG simulation gets closer to observational estimates as reported by Gettelman et al. (2013).

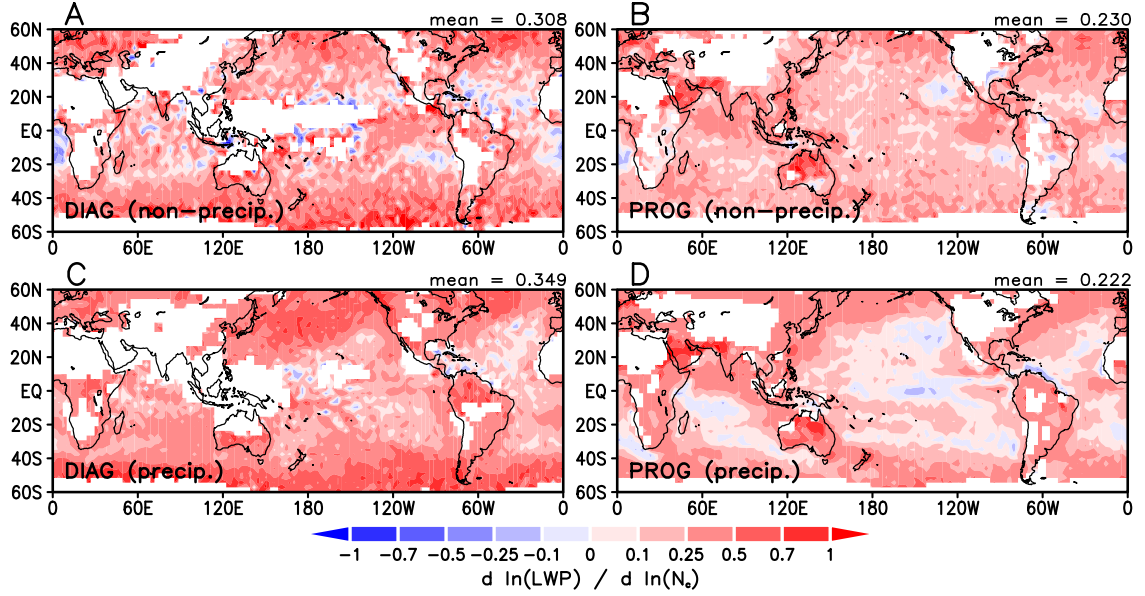


Fig. 6.1: Global distribution of $d \ln LWP / d \ln N_c$ based on (A,C) diagnostic and (B,D) prognostic rain schemes for non-precipitating and precipitating clouds. The threshold of the large-scale precipitation rate of 0.14 mm day^{-1} is used to distinguish between non-precipitating or precipitating conditions. Global mean (60°S – 60°N) of the LWP susceptibility is denoted at the top right in each panel.

The improvement by the PROG described above would result from the enhancement of accretion process over the autoconversion. Figure 6.2 illustrates the joint probability density function (PDF) of LWP and process rates for the DIAG and PROG schemes. In the DIAG scheme, the interaction between cloud droplets and raindrops (i.e., accretion process) is insignificant because the rainwater does not stay in the atmosphere across the multiple time steps, hence the precipitation depends strongly on the autoconversion process (Figs. 6.2A and 6.2). On the other hand, the PROG can memory the rainwater in the atmosphere to the next time step, which enhances the accretion process relative to the autoconversion (Figs. 6.2C and 6.2D). This is consistent with the other reports which tested the prognostic treatment of rain (Posselt and Lohmann, 2008; Gettelman et al., 2015). As a result, the cloud lifetime effect is weakened in the PROG simulation because the accretion process depends only on the masses of cloud water q_c and rainwater q_r whereas the autoconversion depends upon N_c hence aerosol forcing. Since the most of GCMs overestimate the magnitude of the ACI (e.g., Quaas et al., 2009), the above characteristics in the PROG scheme is expected to improve the model biases in the ACI.

It is noteworthy that the PROG shows not only weaker ACI but also the negative response of the LWP susceptibility. This is evident over the ITCZ and over the Indian Ocean in the

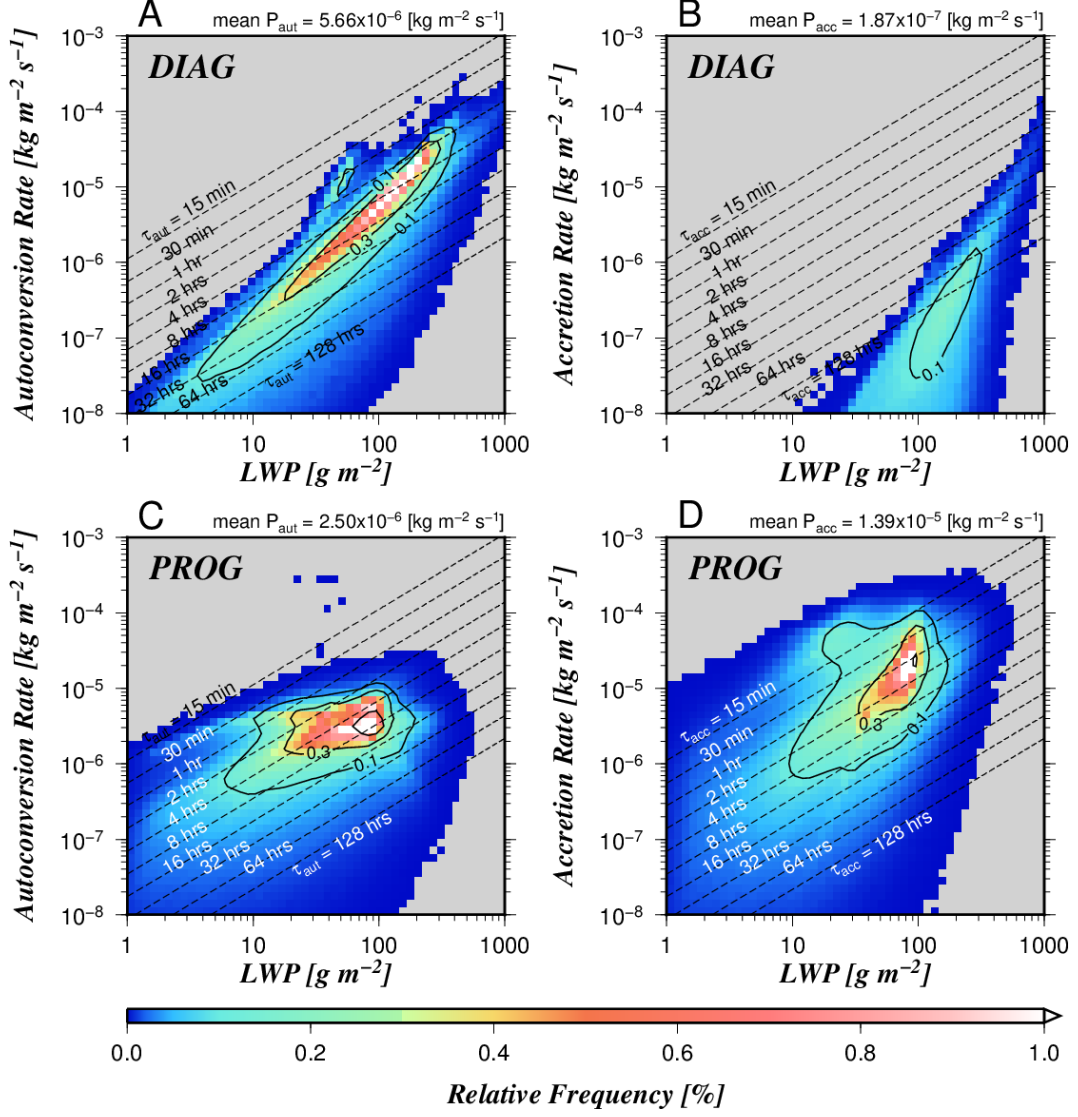


Fig. 6.2: Joint probability density function of LWP and process rates for (A,B) diagnostic and (C,D) prognostic rain schemes. The dashed lines indicate the conversion timescale with regard to the autoconversion and accretion processes as references. Global mean (60°S – 60°N) of the process rates P_{aut} and P_{acc} are denoted at the top right in each panel.

precipitation condition (Fig. 6.1D), whose geographical distribution is analogous to the A-Train estimates (Michibata et al., 2016) which showed a domination of negative LWP susceptibility over the tropics/subtropics. We have confirmed that the characteristic could be found even if other aerosol proxy (hygroscopic aerosol number concentration) was applied instead of N_c , and even if the other autoconversion schemes were employed (not shown). The signal of the negative LWP response appeared in the PROG can be interpreted as one of the buffering behaviors (Stevens and Feingold, 2009), which is supported from the process modeling (e.g., Bretherton et al., 2007;

Lebo and Feingold, 2014; Seifert et al., 2015) using large-eddy simulations (LES).

To represent the “bidirectional” LWP responses is quite difficult in GCMs with any dynamic regimes (Zhang et al., 2016), and overestimate of the ACI particularly cloud lifetime effect is common problem in many of GCMs (e.g., Quaas et al., 2009; Wang et al., 2012; Michibata et al., 2016). The PROG framework in GCMs will be therefore one of the desirable improvements to better representation of the ACI.

6.3.2 Constraining the LWP Susceptibility by Regimes of Cloud Life Stage

We must explain the mechanism of the negative LWP susceptibility appeared in the PROG scheme because the modeled autoconversion parameterization only assumes increase of cloud water with increasing aerosols due to the decrease of precipitation efficiency to represent the cloud lifetime effect. Then the two key questions arise:

1. *What contributes to the negative LWP susceptibility in the PROG scheme?*
2. *How will be the geographical distribution determined linking with the prognostic treatment of rain?*

The most important difference between PROG and DIAG is whether the cloud water and rainwater coexist in the atmospheres or not. Here we examine the regional dependence of the LWP susceptibility on the ratio of the rain water path (RWP) to the total liquid water path (TWP = LWP + RWP) defined as “RWP/TWP”, that is, efficiency of the hydrological cycle. Figure 6.3 shows the LWP susceptibility as a function of RWP/TWP for different 13 regions using 4 autoconversion schemes. Over the three tropics regions (Indian Ocean, ITCZ, and SPCZ) where characterized as a fast-hydrologic cycle show high value of RWP/TWP with low LWP susceptibility. On the other hand, regions of high aerosol loading (Eastern Asia and Japan), calm stratocumulus decks (Peruvian and Californian), and Southern Ocean where characterized as a slower water conversion (Sorooshian et al., 2013) show lower RWP/TWP with higher LWP susceptibility. These geographical characteristics exhibit remarkably good agreement with Michibata et al. (2016) even the magnitude of LWP susceptibility is still different among the model and A-Train observations.

As shown in Fig. 6.3, the simulated LWP susceptibility correlates with the RWP/TWP, and its dependence (i.e., slope) is peculiar to the autoconversion scheme. This would be a result from the different nonlinear dependence of autoconversion timescale τ_{aut} upon N_c among the schemes,

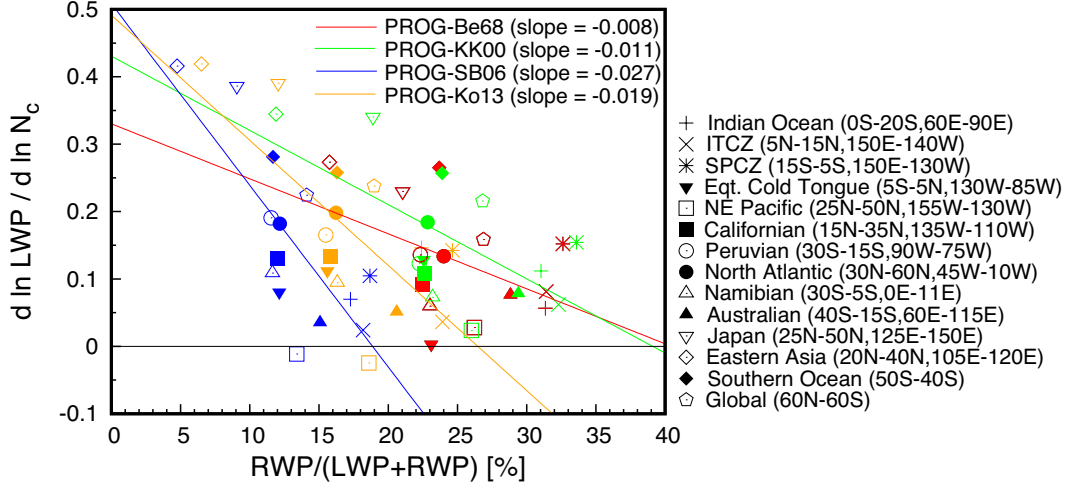


Fig. 6.3: LWP susceptibility as a function of the ratio of rain water path to total liquid water path (RWP/TWP) for different regions based on the PROG simulation. The four autoconversion schemes (Berry (1968) in red; Khairoutdinov and Kogan (2000) in green; Seifert and Beheng (2001) in blue; and Kogan (2013) in orange) are examined.

which controls the diversity (variation) of the precipitation intensity (Suzuki et al., 2013a, 2015); the schemes of Seifert and Beheng (2001) ($\tau_{aut} \propto N_c^{2.0}$) and Kogan (2013) ($\tau_{aut} \propto N_c^{3.01}$) show steeper slope of the regression than those of Berry (1968) ($\tau_{aut} \propto N_c^{1.0}$) and Khairoutdinov and Kogan (2000) ($\tau_{aut} \propto N_c^{1.79}$). In other word, this implies that the improvement of the ACI requires improvements in precipitation intensity and/or frequency as discussed in Figs. 6.11 and 6.2, which support the previous studies (e.g., Wang et al., 2012; Gettelman et al., 2015; Mülmenstädt et al., 2015; Suzuki et al., 2015).

To better understanding the bidirectional LWP response to aerosol perturbations, we focus on the cloud life stage by sorting the regimes into the RWP/TWP and cloud geometric thickness as shown in Fig. 6.4. The positive and negative responses of LWP are systematically classified on the susceptibility matrix. The bottom left region in the diagram, very thin with little ratio of rainwater shows positive LWP susceptibility because the cloud is at the initial condensation stage. The top left region, very thick but less rainwater ratio, exhibits strong positive LWP response with increased aerosols, where the cloud lifetime effect will be evident. The top right region where thick and more rainwater would correspond to mature stage of clouds, and then shows negative LWP susceptibility towards the bottom right region where is accompanied by cloud thinning with dominant rainwater (i.e., decaying stage).

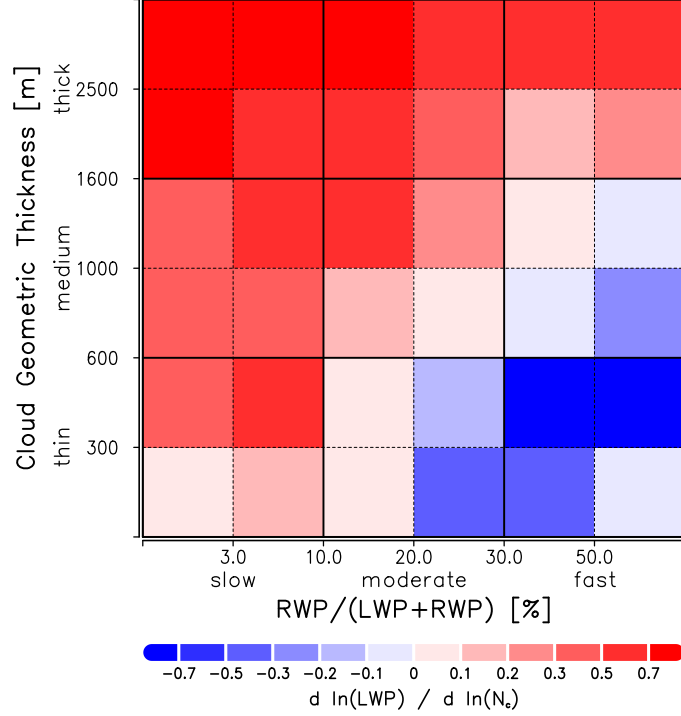


Fig. 6.4: Susceptibility matrix of the LWP response to N_c as a function of RWP/TWP and cloud geometric thickness based on the PROG simulation.

As shown in Fig. 6.4, the bidirectional LWP responses depend strongly on the cloud life stage, and the fundamental mechanism of negative LWP susceptibility would be caused by the high ratio of RWP to TWP and thus acceleration of the accretion, which cannot be represented in the DIAG scheme. This interpretation is supported by sensitivity experiments so as not to enhance the accretion rate P_{acc} against to the autoconversion rate P_{aut} (i.e., $P_{acc} = \min(P_{aut}, P_{acc})$, thus $P_{acc}/P_{aut} < 1$), which show only the positive LWP susceptibility across the globe (not shown). The increased aerosols work in the following two opposite ways, which results in 1) increase of LWP due to the decreased autoconversion rate so-called the cloud lifetime effect, and 2) decrease of LWP due to enhanced accretion in the next time step caused by the coexisted drizzle in the cloud layers.

Guo et al. (2015) also succeeded to represent the LWP responses both positive and negative to aerosol perturbations using macrophysics module “CLUBB”. They suggested that the negative LWP response depends on relative humidity above the boundary layer related to entrainment and evaporation, while our model cannot resolve such the interactions with dynamics (not shown) because incorporated PROG scheme in this study does not coupled to dynamics module currently. The microphysics-dynamics interaction such the entrainment-evaporation feedback (Small et al.,

2009; Seifert et al., 2015) will be also indispensable to incorporate buffering behaviors into GCMs toward more realistic work suggested by process modeling.

6.4 Summary Perspectives

To better understand how treatment of rain in GCMs improve the regime dependence of ACI for warm clouds, a prognostic rain scheme (PROG) was employed in the MIROC-SPRINTARS climate model and compared with the previous diagnostic rain scheme (DIAG). Given that the PROG scheme preserves a “memory” of rainwater during the course of model integration, the drizzle and raindrop thus kept in the atmosphere operate to enhance contributions of the accretion process over those of the autoconversion process. Since only the latter process is a pathway through which the aerosols influence the rain formation, enhanced contributions of accretion lead to reduced magnitudes of the ACI.

This modulation of the rain formation process was also shown to help mitigate the overestimates of the cloud water susceptibility to aerosol perturbations in traditional climate models. Particularly notable is that both increasing and decreasing responses of LWP to increased aerosols were reproduced with the PROG scheme in better agreement with characteristics obtained from satellite observations and process modeling in the literature. This is in stark contrast to what was obtained from the traditional DIAG schemes that reproduced the positive LWP response alone as the so-called cloud lifetime effect. The “bidirectional” LWP responses simulated by the model with the PROG were further found to appear in relation to the life stage of the cloud system that tended to be better represented by the prognostic treatment of rain. As a result, the PROG scheme reduced the LWP susceptibility with a magnitude of -30% against to the DIAG scheme. These improvements propose that the PROG scheme is promising for better process representations of not only cloud-precipitation microphysics and cloud radiative forcing, but also cloud system responses to aerosol perturbations including the “buffered system” behaviors.

The PROG scheme tested in this study does not coupled to the feedback onto environment. Since the entrainment-sedimentation-evaporation feedback (Bretherton et al., 2007; Wood, 2012; Lebo and Feingold, 2014) is one of the indispensable processes to consider the buffering of the aerosol forcing, a further study so as to incorporate such the microphysics-dynamics interactions into GCMs should be conducted. The PROG scheme may be applicable to include their system because evaporation of drizzle and raindrops during the sedimentation can be represented within a short sub-time step (cf. Model and Methods). To include the feedback of microphysics

onto boundary layer stabilization/destabilization due to drizzling stratocumulus based on PROG scheme will be a significant challenge in next generation of GCMs for more accurate climate simulation.

7 General Conclusion

7.1 Contributions and Applications

In this thesis, aerosol-cloud-precipitation-climate interactions were discussed for better understanding of the whole climate system, using satellite data and a global aerosol-climate model (MIROC-SPRINTARS). The main focus of satellite analysis aimed at revealing the source of discrepancies between observations and climate modeling. Here we proposed the several unique methods to understand the complex interactions at fundamental process-levels, which provided indispensable information on the model improvement. The state-of-the-art two-moment bulk microphysics scheme including prognostic precipitation was developed, and successfully implemented into the MIROC-SPRINTARS GCM. The contributions and its applications provided by this work are detailed below.

In Chapter 2, warm rain microphysics over the region from East Asia to the North Pacific in the mid-latitudes was investigated with the joint use of CloudSat/CPR and Aqua/MODIS retrievals. The systematic differences in cloud micro- and macrophysics between land and ocean, as well as those between JJA and DJF were effectively illustrated by “contoured frequency by optical depth diagram (CFODD)” which shows vertical microphysical structure (e.g., Suzuki et al., 2010; Nakajima et al., 2010). The transition of CFODD from cloud, drizzle, to precipitation indicated that the suppression of drizzle/precipitation due to high aerosols concentration mainly occurred around the cloud base over polluted land areas in the stable environment. Although the drizzle occurrence varied from 20 % to 50 % depending on aerosol loading and is consistent with in-situ and satellite observations (e.g., Leon et al., 2008; Kubar et al., 2009; Wu et al., 2015), it was synchronized to the surrounding environments as well. These results suggest that the microphysics scheme assumed in GCMs which consider only microphysical aspects but not dynamics is missing the intrinsic triggers of microphysics-dynamics interactions.

In Chapter 3, the performance of five autoconversion (mass transfer from cloud water to rainwater) schemes in warm rain was evaluated to examine the potential biases in aerosol-cloud interactions. The autoconversion scheme proposed by Tripoli and Cotton (1980) tended to diagnose the highest autoconversion rate resulting in too large precipitation rate. Berry (1968) scheme which is the default scheme in the MIROC-SPRINTARS also demonstrated higher autoconversion rate. Although the scheme proposed by Liu and Daum (2004) showed the lowest precipitation generation among the five schemes, this scheme significantly underestimated the shortwave cloud

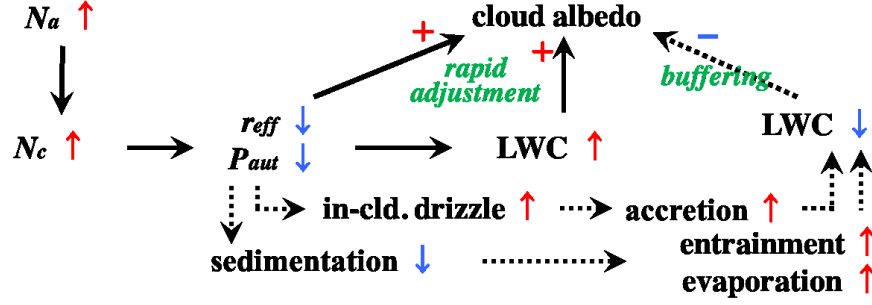


Fig. 7.1: Schematic diagram of cloud system responses to aerosol perturbations revealed in this thesis. The traditional GCMs with diagnostic treatment of rain represent only the positive response of cloud albedo to increased aerosols as shown in solid arrows (described in Chapter 4). The prognostic precipitation framework developed in Chapter 5 can reproduce both positive and negative responses of liquid water, which “buffer” the aerosol perturbations as denoted in the dotted arrows (described in Chapter 6).

radiative forcing (SWCRF) more than -10 W m^{-2} relative to the CERES satellite observation, due to the most remnant of the cloud water. Consequently, the so-called “too few too bright low-cloud problem” (Nam et al., 2012) and “too fast too frequent rain formation” (Suzuki et al., 2013a, 2015) which are the common issue in GCMs were not improved in any configurations. This strongly suggests that the fundamental errors still exist in not only microphysics but also other processes in the model.

In Chapter 4, responses of cloud system (micro- and macrophysics) to aerosol perturbations were explored via a process-oriented model evaluation, to reveal the source of discrepancies in the aerosol-cloud-precipitation interactions between GCMs and A-Train retrievals. GCMs tended to increase cloud water for increased aerosols because the models decrease the autoconversion rate to suppress generation of the rainwater assuming the cloud lifetime effect (Albrecht, 1989). On the other hand, satellite retrievals exhibited both increasing and decreasing responses in cloud water to aerosol perturbations, which was in stark contrast to the model. This “bidirectional” cloud response was closely related to the environmental conditions (e.g., cloud types, thermodynamic condition, and drizzle intensity). The suppression of precipitation by increased aerosols can also alter the in-cloud turbulence kinetic energy (TKE) due to the lower sedimentation flux, which further causes positive feedback of entrainment-evaporation enhancement (Small et al., 2009; Wood, 2012). This “microphysics-dynamics interactions” (Michibata et al., 2016) must be indispensable to reproduce the bidirectional cloud response, which buffers the aerosol perturbations (Stevens and Feingold, 2009) as shown in Fig. 7.1.

In Chapter 5, physically based two-moment bulk microphysics was developed and implemented into MIROC-SPRINTARS, to overcome the fundamental flaw of GCMs suggested in the previous Chapters. The prognostic treatment of rain (PROG) is a special feature of the new scheme. An additional hydrometeor class of rainwater significantly increased the relative contribution of accretion against to the autoconversion, in good agreement with in-situ aircraft measurements (Wood, 2005b; Gettelman et al., 2013). This is because the PROG scheme can keep the memory of rainwater in the atmosphere across the multiple time steps, which also contributed to represent the horizontal distribution of drizzling stratocumuli. As a result, the common bias among GCMs, “too few too bright low-cloud problem” (Nam et al., 2012) was significantly reduced in the PROG simulation.

Furthermore, the ACI was also significantly improved in the new PROG scheme, as discussed in Chapter 6. LWP susceptibility was reduced approximately 30% by applying the PROG scheme, whereas the traditional DIAG approach overestimated the ACI relative to observations (e.g., Quaas et al., 2009; Wang et al., 2012). It is noteworthy that the PROG could represent the bidirectional LWP susceptibility as suggested from A-Train satellite described in Chapter 4. These improvements found in the new scheme propose that the PROG scheme will be promising for better process representations of not only cloud-precipitation microphysics and cloud radiative forcing, but also the response of cloud system to aerosols with the intrinsic “buffered system” morphology.

7.2 Future Work

Although many advantages were found in the PROG, some deteriorations were also appeared (e.g., vertical distribution of cloud liquid and ice water, too few cloud ice). The cloud-precipitation microphysics must be elaborated at a fundamental process-level, and requires more sophisticated framework. This will be addressed in future studies described below.

The roadmap is shown in Fig. 7.2. The primary goal of the future project is to add further two prognostic equations for snow mass and number mixing ratios. The prognostic treatment of both liquid and solid precipitation will be a more physically based representation of interactions among hydrometeors. More specifically, depositional growth of ice crystals and snowflakes from cloud droplets (Bergeron-Findeisen process) is quite important process to reproduce the realistic microphysics in the atmosphere (Cesana et al., 2015). Although some GCMs consider this process, it is difficult for GCMs with long time step to simulate accurately due to its significantly large

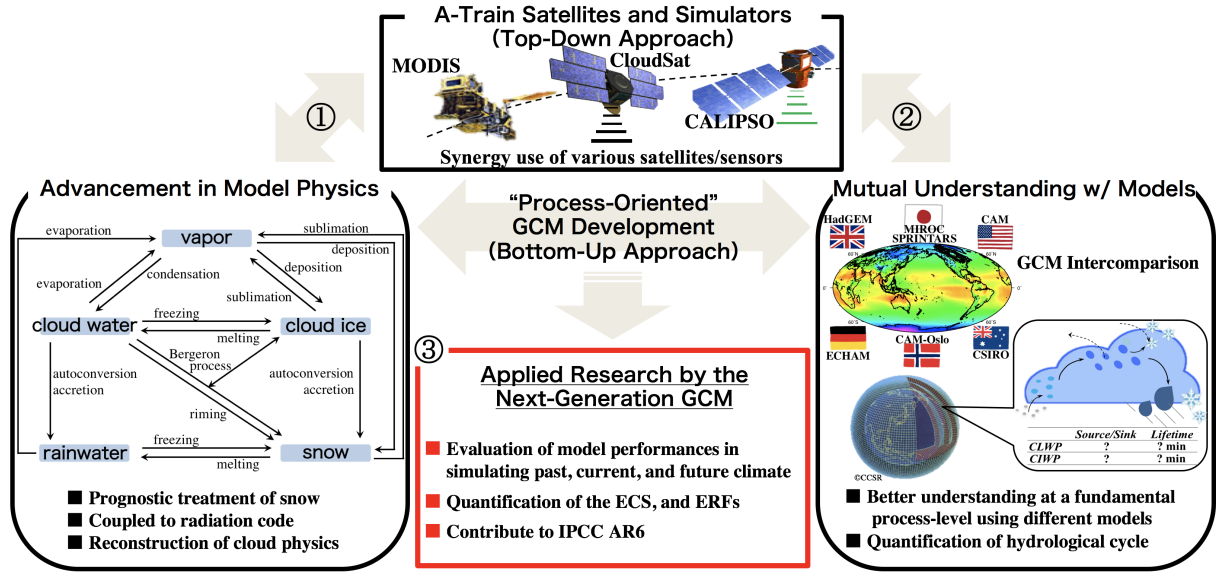


Fig. 7.2: Schematic illustration for construction of the next-generation climate model using satellite observations, satellite simulators, and process modeling. The model will be evaluated in a model intercomparison project (MIP) via the process-oriented model evaluation (see text for details).

tendency (e.g., Gettelman et al., 2015). Thus GCMs still suffer from representations of the balance of cloud liquid water and cloud ice as well as vertical structure of mixed-phase cloud, which further cause discrepancies in the radiative forcing (Matus and L’Ecuyer, 2017; Cesana and Storelmo, 2017) via cloud-climate feedbacks. Introduction of the prognostic snow will be therefore desirable framework because it can consider the various interactions among other prognostic hydrometeors (i.e., water vapor, liquid cloud water, cloud ice, rainwater) within a short timescale by using sub-time step.

A secondary goal of the future project is to evaluate the new microphysics scheme with other models and satellites via instrument simulators (COSP, Bodas-Salcedo et al., 2011). Although the previous studies diagnosed cloud macrophysical structures or evaluated the mean states of the model using COSP, few studies have attempted to diagnose the microphysics via a process-oriented model evaluations (Suzuki et al., 2015). For example, new methods/metrics with the joint use of several microphysical parameters, such the CFODD and LWP susceptibility ($d \ln LWP / d \ln N_c$) would effectively depict the intrinsic model biases (Chapters 2 and 4).

The materials of the model evaluation include not only GCMs, but also high-resolution process models such as LES and cloud resolving model (CRM). GCMs have been evaluated using similar several models (i.e., GCMs intercomparison project; “MIP”), but the discussion also has

been closed within the framework of GCMs. Since the LES and CRM resolve microphysics and interaction with dynamics more explicitly (Zhou and Penner, 2017), the combined use of the process models must be powerful tools to understand the bias in GCMs, which will provide new scientific perspectives beyond the difference of model resolution and/or tradition.

Future work will be conducted based on the above two concepts as shown in Fig. 7.2; the physically based model development (bottom-up approach) and process-oriented model evaluation with satellites and simulators (top-down approach). In order to estimate the effective radiative forcing (ERF) and equilibrium climate sensitivity (ECS), long-term simulations using the new model must be carried out. Unveiling the link between process-level model improvement and the macroscopic climate indices (e.g., ERF and ECS) would provide further understanding of the model biases, and also contribute to elaborating the model physics for more accurate climate simulation.

8 Data and Code Availability

MODIS Level 3 products are available through the LAADS website¹. SSMI/SSMIS instruments data are obtained from the Remote Sensing Systems² data services. ISCCP D2 datasets are also available online³. GPCP version 2.2 can be obtained from NASA Goddard Space Flight Center⁴, and NOAA-18 retrieved OLR from NOAA Earth System Research Laboratory⁵. CERES EBAF-TOA Edition 2.8 products are available from the NASA Langley Research Center⁶. The original CloudSat data products were provided by the CloudSat Data Processing Center at CIRA/Colorado State University⁷.

The code of box-model for process rate equations implemented in the prognostic precipitation scheme described in Chapter 5 is attached in the Appendix A. The CloudSat and MIROC-SPRINTARS climatological data for single-layer warm cloud to reproduce the all figures used in this thesis can be obtained from the author, Takuro Michibata (michibata@riam.kyushu-u.ac.jp) upon request.

¹<https://ladsweb.nascom.nasa.gov>

²<http://www.remss.com/missions/ssmi>

³<http://isccp.giss.nasa.gov/products/browsed2.html>

⁴<http://precip.gsfc.nasa.gov>

⁵http://www.esrl.noaa.gov/psd/data/gridded/data.interp_OLR.html

⁶http://ceres.larc.nasa.gov/order_data.php

⁷<http://www.cloudsat.cira.colostate.edu>

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Appendix A: Code of Box Model for Microphysics Process Rates

In Chapter 5, a simplified zero-dimensional box model was employed to implement the prognostic precipitation scheme. Here the code of the box model is attached below, which reproduces the tendencies shown in Fig. 5.3. The formulas of process rates in the new cloud-precipitation scheme is embedded in this box model. Physical constants and threshold parameters in the microphysics scheme are summarized in Table A1.

Table A1: A list of physical constants and thresholds parameters assumed in microphysics scheme in the box model.

Parameter	Value	Unit	Definition
RAIR	287.04D0	$\text{J kg}^{-1} \text{K}^{-1}$	gas constant of air
RVAP	1004.D0	$\text{J kg}^{-1} \text{K}^{-1}$	gas constant for water vapor
CP	461.D0	$\text{J kg}^{-1} \text{K}^{-1}$	specific heat at constant pressure for dry air
TMELT	273.15D0	K	melting point of water
EL	2.501D6	J kg^{-1}	latent heats of vaporization
EMELT	3.337D5	J kg^{-1}	latent heats of freezing
EL+EMELT		J kg^{-1}	latent heats of sublimation
RHOW	1000.D0	kg m^{-3}	density of water
RHOI	500.D0	kg m^{-3}	density of ice crystal
RHOS	250.D0	kg m^{-3}	density of snow
PI	3.14159265358979		
EPSV	RAIR/RVAP = 0.62265	none	molec. weight water/air
ES0	611.D0	Pa	saturate e 0deg C
NEWRM	6.55D-11	kg	generated rain mass via autoconversion
CSIAUT	150.D-6	m	critical size of ice crystals for autoconversion of ice to snow
TSAUTI	180.D0	sec	timescale for autoconversion of ice to snow
MI0	2.09D-12	kg	mass of new crystal due to aerosol freezing and growth

```

1  !=====
2  !!  <<  Process Rates in Cloud/Precipitation Microphysics  >>
3  !!
4  !!
5  !!  HISTORY:
6  !!      *  2017.06.12: Zero-dim Box Model for MIROC
7  !!                      with Prognostic Rain+Snow Scheme
8  !!      *  2017.06.13: add CloudSat Analytic Simulator
9  !!      *  2017.08.04: Disassembled into modules
10 !!
11 !!
12 !!  AUTHOR:
13 !!      *  Takuro Michibata (michibata@riam.kyushu-u.ac.jp)
14 !!
15 !!
16 !=====
17
18 PROGRAM PROCRATES
19 USE rate_formulas
20
21 IMPLICIT NONE
22
23
24 !--- MODEL VARIABLES:
25 INTEGER    ::  IJ,  K                !! lon/lat/alt index
26 REAL*8     ::  GDT,  GDTX            !! temperature [K]
27 REAL*8     ::  GDP                  !! pressure [Pa]
28 REAL*8     ::  RHO                  !! air density [kg m-3]
29 LOGICAL    ::  ODEBUG                !! debug option
30
31 !--- HYDROMETEORS:
32 REAL*8     ::  RH                    !! relative humidity
33 REAL*8     ::  QSAT                  !! saturation humidity
34 REAL*8     ::  GDCFRC,  PREFRC       !! subgrid fraction
35 REAL*8     ::  GDCFRCI,  SNWFRC      !! subgrid fraction
36 REAL*8     ::  GDQX,  GDQXI          !! H2O vapor mixing ratio
37                                     [kg/kg]
38 REAL*8     ::  GDQVL,  GDQVI         !! H2O vapor mixing ratio
39                                     [kg/kg]
40 REAL*8     ::  GDQC,  GDQI,  GDQR,  GDQS  !! mass mixing ratio [
41                                     kg kg-1]
42 REAL*8     ::  GDNC,  GDNI,  GDNr,  GDNS  !! number concentration [m
43                                     -3]
44 REAL*8     ::  QCLD,  QICE,  QRAIN,  QSNOW  !! mass mixing ratio (in-
45                                     cloud) [kg kg-1]
46 REAL*8     ::  NCLD,  NICE,  NRAIN,  NSNOW  !! number concentration (
47                                     in-cloud) [m-3]
48 REAL*8     ::  QCTEND,  QITEND,  QRTEND,  QSTEND  !! Q tendency term
49 REAL*8     ::  NCTEND,  NITEND,  NRTEND,  NSTEND  !! N tendency term
50
51 !--- SIZE DISTRIBUTION PARAMETERS:
52 REAL*8     ::  LAMC                  !! slope of the cloud
53                                     droplet distribution [m-1]
54 REAL*8     ::  LAMR                  !! slope of the raindrop
55                                     distribution [m-1]
56 REAL*8     ::  LAMI                  !! slope of the cloud ice
57                                     distribution [m-1]
58 REAL*8     ::  LAMS                  !! slope of the snowflake
59                                     distribution [m-1]

```

```

50      !! for DSD
51      REAL*8      :: NUC                !! shape factor (cloud)
52      REAL*8      :: NUR                !! shape factor (rain)
53      REAL*8      :: NUS                !! shape factor (snow)
54      REAL*8      :: MUC                !! shape factor (cloud)
55      REAL*8      :: MUI                !! shape factor (ice)
56      REAL*8      :: MUR                !! shape factor (rain)
57      REAL*8      :: MUS                !! shape factor (snow)
58      REAL*8      :: DCMEAN             !! mean diameter (cloud)
59      [m]
60      REAL*8      :: DIMEAN             !! mean diameter (ice)
61      [m]
62      REAL*8      :: DRMEAN             !! mean diameter (rain)
63      [m]
64      REAL*8      :: DSMEAN             !! mean diameter (snow)
65      [m]
66      REAL*8      :: MASSI              !! ice mass [kg]
67      REAL*8      :: MASSR              !! rain mass [kg]
68      REAL*8      :: MASSS              !! snow mass [kg]
69      REAL*8      :: DIAMI              !! ice crystal diameter [m]
70      ]
71      REAL*8      :: DIAMR              !! raindrop diameter [m]
72      ]
73      REAL*8      :: DIAMS              !! snowflake diameter [um]
74      ]
75      !! sedimentation for rain: Posselt and Lohmann (2009) fallspeed
76      parameter
77      REAL*8      :: DOC                !! DOC = 1/LAMC [m]
78      ]
79      REAL*8      :: DOI                !! DOI = 1/LAMI [m]
80      ]
81      REAL*8      :: DOR                !! DOR = 1/LAMR [m]
82      ]
83      REAL*8      :: DOS                !! DOS = 1/LAMS [m]
84      ]
85      REAL*8      :: VSR                !! terminal velocity [m/s]
86      ]
87      REAL*8      :: VQR                !! mass weight raindrop
88      fall velocity [m/s]
89      REAL*8      :: VNR                !! number weight raindrop
90      fall velocity [m/s]
91      REAL*8      :: VSS                !! terminal velocity [m/s]
92      REAL*8      :: VQS                !! mass weight snow fall
93      velocity [m/s]
94      REAL*8      :: VNS                !! number weight snow fall
95      velocity [m/s]
96      REAL*8      :: DVQ                !! threshold of bulk
97      velocity for VQR [m]
98      REAL*8      :: DVN                !! threshold of bulk
99      velocity for VNR [m]
100     REAL*8      :: BETA1              !! beta1 = 1.23
101     REAL*8      :: BETA2              !! beta2 = 0.22
102     REAL*8      :: NOI                !! intercept: NOI = NICE *
103     LAMI [m-4]
104     REAL*8      :: NOR                !! intercept: NOR = NRAIN*
105     LAMR [m-4]
106     REAL*8      :: NOS                !! intercept: NOS = NSNOW*
107     LAMS [m-4]
108     REAL*8      :: RFLX                !! rain flux [kg m-2 s-1]
109     REAL*8      :: SFLX                !! snow flux [kg m-2 s-1]
110

```

```

89  !--- CloudSat 94GHz Analytic Radar Simulator
90  LOGICAL  ::  OCLSAT                      !! ON/OFF CloudSat
      Simulator
91  REAL*8   ::  CSRR                        !! CloudSat 94GHz radar
      reflectivity in-precip. [dBZe]
92  REAL*8   ::  ACSRR                      !! CloudSat CSRR average
      over grid mean
93  REAL*8   ::  AREFL                      !! average reflectivity [
      dBZe] will zero points outside valid range
94  REAL*8   ::  AREFLZ                    !! average reflectivity [
      Ze].
95
96  !--- DUMMY PARAMETERS:
97  REAL*8   ::  DUM1
98  REAL*8   ::  DUM2
99  REAL*8   ::  DUM3
100
101  !--- PROCESS RATES [kg/kg/s] or [/m3/s]:
102  !! AUTOCONVERSION (LIQUID): C + C -> R
103  REAL*8   ::  PQAUTL                    !! Qr tendency
104  REAL*8   ::  PNAUTLC                   !! Nc tendency
105  REAL*8   ::  PNAUTLR                   !! Nr tendency
106  !! ACCRETION (LIQUID): C + R -> R
107  REAL*8   ::  PQACCL                    !! Qc and/or Qr tendency
108  REAL*8   ::  PNACCL                    !! Nc tendency
109  !! SELF-COLLECTION OF RAIN (LIQUID): R -> R
110  REAL*8   ::  PNSCLL                    !! Nr tendency
111
112  !! AUTOCONVERSION (ICE): I + I -> S
113  LOGICAL  ::  OAUTI
114  REAL*8   ::  PQAUTI
115  REAL*8   ::  PNAUTI
116  !! ACCRETION (ICE): C + S -> S
117  LOGICAL  ::  OACCCS
118  REAL*8   ::  PQACCCS
119  REAL*8   ::  PNACCCS
120  !! (OPTIONAL) HALLET-MOSSOP: S -> I
121  LOGICAL  ::  OHLTMSP
122  REAL*8   ::  PQHLMI
123  REAL*8   ::  PNHLMI
124  !! ACCRETION (ICE): I + S -> S
125  LOGICAL  ::  OACCIS
126  REAL*8   ::  PQACCIS
127  REAL*8   ::  PNACCIS
128  !! ACCRETION (ICE): R + S -> S
129  LOGICAL  ::  OACCRS
130  REAL*8   ::  PQACCRS
131  REAL*8   ::  PNACCRS
132  !! SELF-AGGREGATION (SNOW): S -> S
133  LOGICAL  ::  OSAGS
134  REAL*8   ::  PNSAGS
135  !! BERGERON (ICE):
136  LOGICAL  ::  OBERGI
137  REAL*8   ::  PBERGI
138  !! BERGERON (SNOW):
139  LOGICAL  ::  OBERGS
140  REAL*8   ::  PBERGS
141
142  !--- PROCESS PARAMETERS (TUNING NOBS):
143  REAL*8   ::  TSAUTI                    !! Time-Scale for Autoconversion of Ice,
      assuming 180 sec.

```

```

144 REAL*8      :: NEWSM      !! generated snow mass [kg] via autoconversion
    , assuming 150 micron
145 REAL*8      :: EFRIM      !! collision EFFiciency for RIMing
146 REAL*8      :: EFACRS     !! collision EFFiciency for ACretion of
    Raindrop by Snowflake
147 REAL*8      :: EFAGG      !! AGGregation EFiciency of cloud ice and snow
148 REAL*8      :: AFR        !! ventilation parameters for rain
149 REAL*8      :: BFR        !! ventilation parameters for rain
150 REAL*8      :: AFS        !! ventilation parameters for snow
151 REAL*8      :: BFS        !! ventilation parameters for snow
152 !! fall speed parameters, V = aD^b (V is in m/s)
153 !! droplets
154 REAL*8      :: BC
155 !! cloud ice
156 REAL*8      :: AI
157 REAL*8      :: BI
158 !! rain
159 ! REAL*8      :: AR
160 ! REAL*8      :: BR
161 !! snow
162 REAL*8      :: AS, ASN
163 REAL*8      :: BS
164
165
166 !--- PERTURBED SIMPLE EXPERIMENTS:
167 LOGICAL :: OEXP !! ON/OFF EXPERIMENT
168 INTEGER :: ITMP
169 INTEGER :: IQX
170 INTEGER :: IQC, INC, IQI, INI
171 INTEGER :: IQR, INR, IQS, INS
172 INTEGER, PARAMETER :: NTMP = 10
173 INTEGER, PARAMETER :: NQX = 3
174 INTEGER, PARAMETER :: NQC = 3
175 INTEGER, PARAMETER :: NNC = 3
176 INTEGER, PARAMETER :: NQI = 3
177 INTEGER, PARAMETER :: NNI = 3
178 INTEGER, PARAMETER :: NQR = 3
179 INTEGER, PARAMETER :: NNR = 3
180 INTEGER, PARAMETER :: NQS = 3
181 INTEGER, PARAMETER :: NNS = 3
182 REAL*8 :: QXGVN( NQX )
183 REAL*8 :: QCGVN( NQC ), NCGVN( NNC )
184 REAL*8 :: QIGVN( NQI ), NIGVN( NNI )
185 REAL*8 :: QRGVN( NQR ), NRGVN( NNR )
186 REAL*8 :: QSGVN( NQS ), NSGVN( NNS )
187
188 !! PRESCRIBED PARAMETERS
189 DATA QXGVN / 1.D-3, 5.D-4, 8.D-5 /
190 DATA QCGVN / 3.D-4, 6.D-5, 1.D-5 /
191 DATA NCGVN / 3.D+8, 1.D+8, 3.D+7 /
192 DATA QIGVN / 3.D-5, 8.D-6, 2.D-6 /
193 DATA NIGVN / 3.D+6, 1.D+6, 5.D+5 /
194 DATA QRGVN / 8.D-5, 4.D-5, 6.D-6 /
195 DATA NRGVN / 3.D+4, 1.D+4, 5.D+3 /
196 DATA QSGVN / 3.D-5, 5.D-6, 1.D-6 /
197 DATA NSGVN / 3.D+4, 1.D+4, 5.D+3 /
198
199
200 ! =====
201 ! CONFIGURATION
202 ! =====

```

```

203
204      !--- USER DEFINED PARAMETER (assumptions):
205      GDP      = 650.D2
206      GDCFRC   = 1.D0
207      PREFRC   = 1.D0
208      TSAUTI   = 180.D0
209      NEWSM    = (4.D0/3.D0 * PI * RHOS) * (150.D-6)**(3.0)
210      EFRIM    = 1.D0
211      EFACRS   = 1.D0
212      EFAGG    = 0.5D0
213      !! if MU* is set to 1.0, DSD is defined as the exponential distribution,
214      !! and other integer 1 < MU* defined as modified-gamma distribution.
215      NUC      = 2.D0
216      NUR      = 0.3333333333333333D0
217      MUC      = 1.D0
218      MUI      = 1.D0
219      MUR      = 5.D0
220      MUS      = 1.D0
221      DVQ      = DVR / ( MUR + 3.D0 )
222      DVN      = DVR / MUR
223      BETA1    = 1.23D0
224      BETA2    = 0.22D0
225      !! fall speed parameters, V = aD^b (V is in m/s)
226      BC = 2.D0          !! droplets
227      AI = 700.D0        !! cloud ice
228      BI = 1.D0          !! cloud ice
229      !   AR = 841.99667D0 !! rain
230      !   BR = 0.8D0       !! rain
231      AS = 11.72D0        !! snow
232      BS = 0.41D0        !! snow
233      !! ventilation parameters (based on Khairoutdinov and Randall, 2003)
234      AFR = 0.78D0        !! for rain
235      BFR = 0.308D0       !! for rain
236      AFS = 0.86D0        !! for snow
237      BFS = 0.28D0        !! for snow
238      OAUTI   = .TRUE.    !! ON/OFF: AUTOCONVERSION OF ICE
239      OACCCS  = .TRUE.    !! ON/OFF: ACCRETION C+S
240      OHLTMSP = .TRUE.    !! ON/OFF: HALLET-MOSSOP
241      OACCIS  = .TRUE.    !! ON/OFF: ACCRETION I+S
242      OACCRS  = .TRUE.    !! ON/OFF: ACCRETION R+S
243      OSAGS   = .TRUE.    !! ON/OFF: SELF-AGGREGATION OF SNOW
244      OBERGI  = .TRUE.    !! ON/OFF: BERGERON FOR ICE
245      OBERGS  = .TRUE.    !! ON/OFF: BERGERON FOR SNOW
246      !!   OCLSAT = .TRUE.  !! ON/OFF: CLOUDSAT ANALYTIC SIMULATOR
247      OCLSAT  = .FALSE.   !! ON/OFF: CLOUDSAT ANALYTIC SIMULATOR
248      !!   ODEBUG = .TRUE.  !! DEBUG OPTION (TRUE --> OUTPUT LOG-MESSAGES)
249      ODEBUG  = .FALSE.   !! DEBUG OPTION (TRUE --> OUTPUT LOG-MESSAGES)
250      OEXP    = .TRUE.
251      !!   OEXP   = .FALSE.
252
253
254      !=====
255      !   MAIN PROCEDURE
256      !=====
257
258      IF( OEXP ) THEN
259          !--- OUTPUT FILE OPEN:
260          !! local (ic-cloud or in-precip) value
261          OPEN( 11, FILE = "../transout/tendency_mass_local.txt", STATUS = "
                UNKNOWN" )

```

```

262     OPEN( 12, FILE = "../transout/tendency_numb_local.txt", STATUS = "
        UNKNOWN" )
263
264     !! grid averaged value
265     OPEN( 21, FILE = "../transout/tendency_mass_grdav.txt", STATUS = "
        UNKNOWN" )
266     OPEN( 22, FILE = "../transout/tendency_numb_grdav.txt", STATUS = "
        UNKNOWN" )
267
268
269
270     !--- EXPERIMENTS:
271     DO ITMP = 1, NTMP
272     DO IQX = 1, NQX
273     DO IQC = 1, NQC
274     DO INC = 1, NNC
275     DO IQI = 1, NQI
276     DO INI = 1, NNI
277     DO IQR = 1, NQR
278     DO INR = 1, NNR
279     DO IQS = 1, NQS
280     DO INS = 1, NNS
281
282     WRITE(*,*) "ITMP, IQX, IQC, INC, IQI, INI, IQR, INR, IQS, INS"
283     WRITE(*,*) ITMP, IQX, IQC, INC, IQI, INI, IQR, INR, IQS, INS
284
285     !! GET THE PRESCRIBED VALUE
286     GDT = 275.DO - REAL( ITMP - 1.DO )
287     GDTX = GDT
288     GDQX = QXGVN( IQX )
289     GDQC = QCGVN( IQC )
290     GDNC = NCGVN( INC )
291     GDQI = QIGVN( IQI )
292     GDNI = NIGVN( INI )
293     GDQR = QRGVN( IQR )
294     GDNR = NRGVN( INR )
295     GDQS = QSGVN( IQS )
296     GDNS = NSGVN( INS )
297
298
299     !--- INITIALIZE:
300     QCTEND = 0.DO; QITEND = 0.DO; QRTEND = 0.DO; QSTEND = 0.DO
301     NCTEND = 0.DO; NITEND = 0.DO; NRTEND = 0.DO; NSTEND = 0.DO
302
303     GDCFCI = GDCFRC * ( MAX(GDQI,QIMIN) / MAX(GDQC+GDQI,QIMIN) )
304     SNWFRC = PREFRC * ( MAX(GDQS,QIMIN) / MAX(GDQR+GDQS,QIMIN) )
305     IF( ODEBUG ) THEN
306         WRITE(*,*) "***GDFCI=", GDCFCI
307         WRITE(*,*) "SNWFRC=", SNWFRC
308     ENDIF
309
310     QCLD = GDQC / GDCFRC !! in-cloud
311     QICE = GDQI / GDCFRC !! in-cloud
312     QRAIN = GDQR / PREFRC !! in-precip
313     QSNOW = GDQS / PREFRC !! in-precip
314     NCLD = GDNC / GDCFRC !! in-cloud
315     NICE = GDNI / GDCFRC !! in-cloud
316     NRAIN = GDNR / PREFRC !! in-precip
317     NSNOW = GDNS / PREFRC !! in-precip
318
319

```

```

320      !--- HUMIDITY etc...
321      RHO    = GDP / (RAIR*GDTX)
322      QSAT   = FQSATL( GDTX, GDP )
323      RH     = MAX( GDQX / QSAT, 0.DO )
324      GDQVL  = FQSATL( GDTX, GDP )
325      IF( GDTX .GE. TMELT ) THEN
326          GDQVI = GDQVL
327      ELSE
328          GDQVI = FQSATI( GDTX, GDP )
329      ENDIF
330      IF( ODEBUG ) THEN
331          WRITE(*,*) "***_QSAT_=", QSAT
332          WRITE(*,*) "____RH____=", RH
333          WRITE(*,*) "***_GDQVL_=", GDQVL
334          WRITE(*,*) "____GDQVI_=", GDQVI
335      ENDIF
336
337
338      !--- SIZE PARAMETERS:
339      IF( QICE .GT. 0.DO ) THEN
340          DIMEAN = ((6.DO*RHO*QICE)/(PI*RHOI*NICE))**(1.0/3.0)
341      ELSE
342          DIMEAN = 0.DO
343      ENDIF
344      IF( QRAIN .GT. 0.DO ) THEN
345          DRMEAN = ((6.DO*RHO*QRAIN)/(PI*RHOW*NRAIN))**(1.0/3.0)
346      ELSE
347          DRMEAN = 0.DO
348      ENDIF
349      IF( QSNOW .GT. 0.DO ) THEN
350          DSMEAN = ((6.DO*RHO*QSNOW)/(PI*RHOS*NSNOW))**(1.0/3.0)
351      ELSE
352          QSNOW  = 1.D-18
353          DSMEAN = 1.D-10
354      ENDIF
355
356      IF( DIMEAN .GT. 0.DO .AND. NICE .GT. NIMIN ) THEN
357          DOI    = DIMEAN/((MUI+2.DO)*(MUI+1.DO)*MUI)**(1.0/3.0)
358          LAMI    = 1.DO / DOI
359          LAMI    = ( (PI*RHO*NICE*GAMM(MUI+4.DO))/(6.DO*QICE*GAMM(MUI+1.DO)) )
360                  *(1.DO/3.DO)
361          MASSI   = QICE * RHO / NICE
362          DIAMI   = 1.DO / LAMI * 1.D6
363          NOI     = NICE * LAMI
364          NOI     = LAMI**(4.0) * 6.DO * RHO * QICE / ( PI * RHOI ) / GAMM(3.DO
365                  +MUI)
366      ELSE
367          DOI    = 0.DO
368          MASSI   = 0.DO
369          DIAMI   = 0.DO
370      ENDIF
371      IF( DRMEAN .GT. 0.DO .AND. NRAIN .GT. NRMIN ) THEN
372          DOR     = DRMEAN/((MUR+2.DO)*(MUR+1.DO)*MUR)**(1.0/3.0)
373          LAMR    = 1.DO / DOR
374          LAMR    = ( (PI*RHO*NRAIN*GAMM(MUR+4.DO))/(6.DO*QRAIN*GAMM(MUR+1.DO))
375                  )**(1.DO/3.DO)
376          MASSR   = QRAIN * RHO / NRAIN
377          DIAMR   = ( MASSR / AR )**( 1.0 / BR )
378          NOR     = NRAIN * LAMR
379          NOR     = LAMR**(4.0) * 6.DO * RHO * QRAIN / ( PI * RHOW ) / GAMM(3.DO+
380                  MUR)

```

```

377 ELSE
378     DOR      = 0.D0
379     MASSR    = 0.D0
380     DIAMR    = 0.D0
381 ENDIF
382 IF( DSMEAN .GT. 0.D0 .AND. NSNOW .GT. NSMIN ) THEN
383     DOS      = DSMEAN/((MUS+2.D0)*(MUS+1.D0)*MUS)**(1.0/3.0)
384     LAMS     = 1.D0 / DOS
385     LAMS     = ( (PI*RHO*NSNOW*GAMM(MUS+4.D0))/(6.D0*QSNOW*GAMM(MUS+1.D0))
386               )**(1.D0/3.D0)
387     MASSS    = QSNOW * RHO / NSNOW
388     DIAMS    = 1.D0 / LAMS * 1.D6
389     NOS      = NSNOW * LAMS
390     NOS      = LAMS**(4.0) * 6.D0 * RHO * QSNOW / ( PI * RHOS ) / GAMM(3.D0
391               +MUS)
392 ELSE
393     DOS      = 0.D0
394     MASSS    = 0.D0
395 ENDIF
396
397 !--- SEDIMENTATION OF RAIN AND SNOW:
398 VSR = B1R - B2R*EXP(-B3R*DIAMR) + ( B2R-B1R ) * EXP( -5.D0*B3R*DIAMR )
399 !! DIAMR=[m]
400 VSS = BETA1 * (DIAMS*1.D-3)**(BETA2) !! note: here D is in unit [mm],
401 not [m] or [um]
402 !! mass(Q) and number(N) weighted bulk fall velocities:
403 IF( DOR .LT. DVQ ) THEN
404     VQR = ( MUR + 3.D0 ) * BVR * DOR
405 ELSEIF( DOR .GE. DVQ ) THEN
406     VQR = B1R
407 ENDIF
408 IF( DOR .LT. DVN ) THEN
409     VNR = MUR * BVR * DOR
410 ELSEIF( DOR .GE. DVN ) THEN
411     VNR = B1R
412 ENDIF
413 IF( DOS*1.D6 .GT. 30.D0 ) THEN
414     VQS = BETA1 * DOS**(BETA2) * ( GAMM(3.D0+MUS+BETA2) / GAMM(3.D0+MUS)
415     )
416     VNS = BETA1 * DOS**(BETA2) * ( GAMM(MUS+BETA2) / GAMM(MUS) )
417 ELSE
418     VQS = 0.D0
419     VNS = 0.D0
420 ENDIF
421
422 !! PRECIPITATION FLUX FOR RAIN AND SNOW
423 RFLX = RHO * QRAIN * VQR !! [kg m-2 s-1]
424 SFLX = RHO * QSNOW * VQS !! [kg m-2 s-1]
425
426 !--- DISPLAY THE INPUTS etc...
427 IF( ODEBUG ) THEN
428     WRITE(*,*) "***_BVR,_,_,_,DVR_,_,_,_,", BVR, DVR
429     WRITE(*,*) "***_GDQC,_,_,_,GDNC_,_,_,_", GDQC, GDNC
430     WRITE(*,*) "_,_,_,_,GDQI,_,_,_,GDNI_,_,_,_", GDQI, GDNI
431     WRITE(*,*) "_,_,_,_,GDQR,_,_,_,GDNR_,_,_,_", GDQR, GDNR
432     WRITE(*,*) "_,_,_,_,GDQS,_,_,_,GDNS_,_,_,_", GDQS, GDNS
433     WRITE(*,*) "***_DIMEAN,_,_,_,DRMEAN,_,_,_,DSMEAN_,_,_,_", DIMEAN, DRMEAN, DSMEAN
434     WRITE(*,*) "_,_,_,_,DOI,_,_,_,DOR,_,_,_,DOS_,_,_,_", DOI, DOR, DOS
435     WRITE(*,*) "_,_,_,_,DIAMI,_,_,_,DIAMR,_,_,_,DIAMS_,_,_,_", DIAMI, DIAMR, DIAMS

```

```

433     WRITE(*,*) "    MASSI,    MASSR,    MASSS    =", MASSI,  MASSR,  MASSS
434     WRITE(*,*) "    LAMI,    LAMR,    LAMS    =", LAMI,  LAMR,  LAMS
435     WRITE(*,*) "    NOI,    NOR,    NOS    =", NOI,  NOR,  NOS
436     WRITE(*,*) "*** VSR,    VSS    =", VSR,  VSS
437     WRITE(*,*) "    VQR,    VQS    =", VQR,  VQS
438     WRITE(*,*) "    VNR,    VNS    =", VNR,  VNS
439     WRITE(*,*) "*** RFLX,    SFLX    =", RFLX,  SFLX
440     WRITE(*,*)
441 ENDIF
442
443
444 !--- PROCESS RATES EQUATIONS:
445 !--- << WARM RAIN MICROPHYSICS >>
446
447 !! AUTOCONVERSION (LIQUID): C + C -> R
448 !! based on Khairoutdinov and Kogan (2000)
449 CALL AUTOCONVERSION_LIQ( &
450     & ODEBUG, & !! in
451     & RHO, & !! in
452     & QCLD, & !! in
453     & NCLD, & !! in
454     & QRAIN, & !! in
455     & PQAUTL, & !! out
456     & PNAUTLC, & !! out
457     & PNAUTLR ) !! out
458
459
460 !! ACCRETION (LIQUID): C + R -> R
461 !! based on Khairoutdinov and Kogan (2000)
462 CALL ACCRETION_LIQ( &
463     & ODEBUG, & !! in
464     & QCLD, & !! in
465     & NCLD, & !! in
466     & QRAIN, & !! in
467     & PQACCL, & !! out
468     & PNACCL ) !! out
469
470
471 !! SELF-COLLECTION (LIQUID): R -> R
472 !! based on Beheng (1994)
473 CALL SELFCOLL_LIQ( &
474     & ODEBUG, & !! in
475     & RHO, & !! in
476     & QRAIN, & !! in
477     & NRAIN, & !! in
478     & PNSCLL ) !! out
479
480
481 !--- << COLD RAIN MICROPHYSICS >>
482
483 !! AUTOCONVERSION (ICE): I + I -> S
484 !! based on Ferrier (1994)
485 CALL AUTOCONVERSION_ICE( &
486     & ODEBUG, & !! in
487     & GDTX, & !! in
488     & QICE, & !! in
489     & LAMI, & !! in
490     & NOI, & !! in
491     & PQAUTI, & !! out
492     & PNAUTI ) !! out
493

```

```

494
495      !! ACCRETION (ICE): C + S -> S
496      !! optional Hallet-Mossop process included here
497      !! based on Seifert and Beheng (2006)
498      CALL ACCRETION_CS2S( &
499          & ODEBUG, & !! in
500          & OHLTMS, & !! in
501          & RHO, & !! in
502          & GDTX, & !! in
503          & QCLD, & !! in
504          & QSNOW, & !! in
505          & NCLD, & !! in
506          & LAMS, & !! in
507          & NOS, & !! in
508          & AS, & !! in
509          & BS, & !! in
510          & EFRIM, & !! in
511          & PQACCS, & !! out
512          & PNACCS, & !! out
513          & PQHLMI, & !! out
514          & PNHLMI ) !! out
515
516
517      !! ACCRETION (ICE): I + S -> S
518      !! based on Seifert and Beheng (2006)
519      CALL ACCRETION_IS2S( &
520          & ODEBUG, & !! in
521          & RHO, & !! in
522          & GDTX, & !! in
523          & QICE, & !! in
524          & QSNOW, & !! in
525          & NICE, & !! in
526          & LAMS, & !! in
527          & NOS, & !! in
528          & AS, & !! in
529          & BS, & !! in
530          & EFAGG, & !! in
531          & PQACCIS, & !! out
532          & PNACCIS ) !! out
533
534
535      !! ACCRETION (ICE): R + S -> S
536      !! based on Reisner et al. (1998)
537      CALL ACCRETION_RS2S( &
538          & ODEBUG, & !! in
539          & RHO, & !! in
540          & GDTX, & !! in
541          & QRAIN, & !! in
542          & QSNOW, & !! in
543          & LAMR, & !! in
544          & LAMS, & !! in
545          & NOR, & !! in
546          & NOS, & !! in
547          & VQR, & !! in
548          & VNR, & !! in
549          & VQS, & !! in
550          & VNS, & !! in
551          & EFACRS, & !! in
552          & PQACCRS, & !! out
553          & PNACCRS ) !! out
554

```

```

555
556      !! SELF-AGGREGATION (SNOW): S -> S
557      !! based on Passarelli (1978)
558      CALL SELFAGG_SNOW( &
559          & ODEBUG, & !! in
560          & RHO, & !! in
561          & GDTX, & !! in
562          & QSNOW, & !! in
563          & NSNOW, & !! in
564          & AS, & !! in
565          & BS, & !! in
566          & EFAGG, & !! in
567          & PNSAGS ) !! out
568
569
570      !! BERGERON PROCESS: C -> I
571      !! based on Rotstayn et al. (2000)
572      CALL BERGERON_ICE( &
573          & ODEBUG, & !! in
574          & OBERGI, & !! in
575          & RHO, & !! in
576          & GDTX, & !! in
577          & GDP, & !! in
578          & GDQVL, & !! in
579          & GDQVI, & !! in
580          & QICE, & !! in
581          & LAMI, & !! in
582          & NOI, & !! in
583          & PBERGI ) !! out
584
585
586      !! BERGERON PROCESS: C -> S
587      !! based on Rotstayn et al. (2000)
588      CALL BERGERON_SNOW( &
589          & ODEBUG, & !! in
590          & OBERGS, & !! in
591          & RHO, & !! in
592          & GDTX, & !! in
593          & GDP, & !! in
594          & GDQVL, & !! in
595          & GDQVI, & !! in
596          & QICE, & !! in
597          & QSNOW, & !! in
598          & LAMS, & !! in
599          & NOS, & !! in
600          & AS, & !! in
601          & BS, & !! in
602          & AFS, & !! in
603          & BFS, & !! in
604          & PBERGS ) !! out
605
606
607      !--- SUMMARY TENDENCY:
608      !! note: PBERGI is already grid-mean value.
609      !! CLOUD WATER !!
610      QCTEND = QCTEND + &
611          & (-PQAUTL-PQACCL-PQHLMI-PQACCS-PBERGS)*GDCFR - &
612          & PBERGI
613      NCTEND = NCTEND + &
614          & (-PNAUTLC-PNACCL-PNACCS)*GDCFR
615

```

```

616      !! CLOUD ICE !!
617      QITEND = QITEND + &
618              & PQHLM1*GDCFRFC + &
619              & (-PQAUT1-PQACCIS)*GDCFRFCI + &
620              & PBERGI
621      NITEND = NITEND + &
622              & PNHLMI*GDCFRFC + &
623              & (-PNAUT1-PNACCIS)*GDCFRFCI
624
625      !! RAINWATER !!
626      QRTEND = QRTEND + &
627              & (PQAUTL+PQACCL)*GDCFRFC + &
628              & (-PQACCRS)*PREFRC
629      NRTEND = NRTEND + &
630              & (PNAUTLR)*GDCFRFC + &
631              & (-PNACCRS+PNSCLL)*PREFRC  !! note: PNSCLL is already negative
632
633      !! SNOWFLAKE !!
634      QSTEND = QSTEND + &
635              & (PQACCS+PBERGS)*GDCFRFC + &
636              & (PQAUT1+PQACCIS)*GDCFRFCI + &
637              & (PQACCRS)*PREFRC
638      NSTEND = NSTEND + &
639              & PNSAGS*SNWFRC + &
640              & PNAUT1*GDCFRFCI  !! note: PNSAGS is already negative
641
642      !! DIAPLAY THE RESULTS
643      IF( ODEBUD ) THEN
644          WRITE(*,*) "*****"
645          WRITE(*,*) "+++++++SUMMARYTENDENCIES+++++++"
646          WRITE(*,*) "***QCTEND= ", QCTEND
647          WRITE(*,*) "NCTEND= ", NCTEND
648          WRITE(*,*) "***QITEND= ", QITEND
649          WRITE(*,*) "NITEND= ", NITEND
650          WRITE(*,*) "***QRTEND= ", QRTEND
651          WRITE(*,*) "NRTEND= ", NRTEND
652          WRITE(*,*) "***QSTEND= ", QSTEND
653          WRITE(*,*) "NSTEND= ", NSTEND
654          WRITE(*,*) "*****"
655      ENDIF
656
657
658      !--- CloudSat Analytic Radar Simulator:
659      IF( OCLSAT ) THEN
660          WRITE(*,*) "*****"
661          WRITE(*,*) "%%CALLINGCLOUDSATSIMULATOR.."
662          CALL CLSAT94_ANLSIM &
663              & ( GDT, GDTX, GDP, & !! in
664                  & GDCFRFC, PREFRC, & !! in
665                  & GDQC, GDQI, GDQR, GDQS, & !! in
666                  & GDNC, GDNI, GDNr, GDNS, & !! in
667                  & RFLX, & !! in
668                  & CSRR, ACSRR, AREFL, AREFLZ ) !! out
669          WRITE(*,*) "***CSRR= ", CSRR
670          WRITE(*,*) "ACSRR= ", ACSRR
671          WRITE(*,*) "AREFL= ", AREFL
672          WRITE(*,*) "AREFLZ= ", AREFLZ
673      ENDIF
674
675
676      !--- OUTPUTs:

```

```

677      !! TENDENCY FOR MASS MIXING RATIOS (Q):
678      WRITE( 11,* ) "1", QCTEND, & !! Qc tend (all)
679      & "2", -PQAUTL, & !! Qc tend
680      & "3", -PQACCL, & !! Qc tend
681      & "4", -PQHLMI, & !! Qc tend
682      & "5", -PQACCS, & !! Qc tend
683      & "6", -PBERGS, & !! Qc tend
684      & "7", -PBERGI, & !! Qc tend
685      & "8", QITEND, & !! Qi tend (all)
686      & "9", PQHLMI, & !! Qi tend
687      & "10", -PQAUTI, & !! Qi tend
688      & "11", -PQACCIS, & !! Qi tend
689      & "12", PBERGI, & !! Qi tend
690      & "13", QRTEND, & !! Qr tend (all)
691      & "14", PQAUTL, & !! Qr tend
692      & "15", PQACCL, & !! Qr tend
693      & "16", -PQACCRS, & !! Qr tend
694      & "17", QSTEND, & !! Qs tend (all)
695      & "18", PQACCS, & !! Qs tend
696      & "19", PBERGS, & !! Qs tend
697      & "20", PQAUTI, & !! Qs tend
698      & "21", PQACCIS, & !! Qs tend
699      & "22", PQACCRS, & !! Qs tend
700
701      !! TENDENCY FOR NUMBER MIXING RATIOS (N):
702      WRITE( 12,* ) "1", NCTEND, & !! Nc tend (all)
703      & "2", -PNAUTLC, & !! Nc tend
704      & "3", -PNACCL, & !! Nc tend
705      & "4", -PNACCS, & !! Nc tend
706      & "5", NITEND, & !! Ni tend (all)
707      & "6", PNHLMI, & !! Ni tend
708      & "7", -PNAUTI, & !! Ni tend
709      & "8", -PNACCIS, & !! Ni tend
710      & "9", NRTEND, & !! Nr tend (all)
711      & "10", PNAUTLR, & !! Nr tend
712      & "11", -PNACCRS, & !! Nr tend
713      & "12", -PNSCLL, & !! Nr tend
714      & "13", NSTEND, & !! Ns tend (all)
715      & "14", PNSAGS, & !! Ns tend
716      & "15", PNAUTI, & !! Ns tend
717
718      !! TENDENCY FOR MASS MIXING RATIOS (Q):
719      WRITE( 21,* ) "1", QCTEND, & !! Qc tend (all)
720      & "2", -PQAUTL*GDCFRC, & !! Qc tend
721      & "3", -PQACCL*GDCFRC, & !! Qc tend
722      & "4", -PQHLMI*GDCFRC, & !! Qc tend
723      & "5", -PQACCS*GDCFRC, & !! Qc tend
724      & "6", -PBERGS*GDCFRC, & !! Qc tend
725      & "7", -PBERGI, & !! Qc tend
726      & "8", QITEND, & !! Qi tend (all)
727      & "9", PQHLMI*GDCFRC, & !! Qi tend
728      & "10", -PQAUTI*GDCFRCI, & !! Qi tend
729      & "11", -PQACCIS*GDCFRCI, & !! Qi tend
730      & "12", PBERGI, & !! Qi tend
731      & "13", QRTEND, & !! Qr tend (all)
732      & "14", PQAUTL*GDCFRC, & !! Qr tend
733      & "15", PQACCL*GDCFRC, & !! Qr tend
734      & "16", -PQACCRS*PREFRC, & !! Qr tend
735      & "17", QSTEND, & !! Qs tend (all)
736      & "18", PQACCS*GDCFRC, & !! Qs tend
737      & "19", PBERGS*GDCFRC, & !! Qs tend

```

```

738      & "20",    PQAUTI*GDCFRCI,    &  !! Qs tend
739      & "21",    PQACCIS*GDCFRCI,    &  !! Qs tend
740      & "22",    PQACCRS*PREFRC      !! Qs tend
741
742      !! TENDENCY FOR NUMBER MIXING RATIOS (N):
743      WRITE( 22,* ) "□1",    NCTEND,    &  !! Nc tend (all)
744      & "□2",    -PNAUTLC*GDCFRFC,    &  !! Nc tend
745      & "□3",    -PNACCL*GDCFRFC,    &  !! Nc tend
746      & "□4",    -PNACCCS*GDCFRFC,    &  !! Nc tend
747      & "□5",    NITEND,    &  !! Ni tend (all)
748      & "□6",    PNHLMI*GDCFRFC,    &  !! Ni tend
749      & "□7",    -PNAUTI*GDCFRCI,    &  !! Ni tend
750      & "□8",    -PNACCIS*GDCFRCI,    &  !! Ni tend
751      & "□9",    NRTEND,    &  !! Nr tend (all)
752      & "10",    PNAUTLR*GDCFRFC,    &  !! Nr tend
753      & "11",    -PNACCRS*PREFRC,    &  !! Nr tend
754      & "12",    -PNSCLL*PREFRC,    &  !! Nr tend
755      & "13",    NSTEND,    &  !! Ns tend (all)
756      & "14",    PNSAGS*SNWFRC,    &  !! Ns tend
757      & "15",    PNAUTI*GDCFRCI      !! Ns tend
758
759
760      ENDDO
761      ENDDO
762      ENDDO
763      ENDDO
764      ENDDO
765      ENDDO
766      ENDDO
767      ENDDO
768      ENDDO
769      ENDDO
770
771
772      CLOSE( 11 ); CLOSE( 12 )
773      CLOSE( 21 ); CLOSE( 22 )
774
775
776      STOP
777      END PROGRAM PROCRAATES
778      ! ===== !

```

```

1
2 !=====
3 !                               F U N C T I O N S                               !
4 !=====
5
6 !==== Gamma Function =====!
7 REAL(8) FUNCTION GAMM( XX )
8
9     REAL(8)    XX
10    INTEGER    J
11    REAL(8)    SER, STP, TMP, X, Y, COF( 6 )
12    SAVE       COF, STP
13    DATA      COF, STP &
14        &      / 76.18009172947146D0, -86.50532032941677D0, &
15        &      24.01409824083091D0, -1.231739572450155D0, &
16        &      1.208650973866179D-3, -5.395239384953D-6, &
17        &      2.5066282746310005D0 /
18
19    X = XX
20    Y = X
21    TMP = X + 5.5D0
22    TMP = ( X + 0.5D0 ) * LOG( TMP ) - TMP
23    SER = 1.000000000190015D0
24    DO J = 1, 6
25        Y = Y + 1.D0
26        SER = SER + COF( J ) / Y
27    ENDDO
28    GAMM = EXP( TMP + LOG( STP * SER / X ) )
29
30    RETURN
31 END FUNCTION GAMM
32
33
34
35 !==== Incomplete Gamma Function =====!
36 !! the method based on Blahak (2010,GMD)
37 REAL(8) FUNCTION IGFUNC( IGFA, IGFX )
38
39     REAL(8) :: IGFA, IGFX
40     !! LUT
41     REAL(8) :: IGFP(6)
42     REAL(8) :: IGFQ(4)
43     REAL(8) :: IGFR(4)
44     REAL(8) :: IGFS(5)
45     DATA IGFP / 9.4368392235D-03, -1.0782666481D-04, &
46         &      -5.8969657295D-06, 2.8939523781D-07, &
47         &      1.0043326298D-01, 5.5637848465D-01 /
48     DATA IGFQ / 1.1464706419D-01, 2.6963429121D+00, &
49         &      -2.9647038257D+00, 2.1080724954D+00 /
50     DATA IGFR / 0.D0, 1.1428716184D+00, &
51         &      -6.6981186438D-03, 1.0480765092D-04 /
52     DATA IGFS / 1.0356711153D+00, 2.3423452308D+00, &
53         &      -3.6174503174D-01, -3.1376557650D+00, &
54         &      2.9092306039D+00 /
55
56     !! Complete Gamma Function with the argument "a"
57     REAL(8) :: CGF
58     !! Approximate Parameters for IGF
59     REAL(8) :: IGFC1
60     REAL(8) :: IGFC2

```

```

60 REAL(8) :: IGFC3
61 REAL(8) :: IGFC4
62 REAL(8) :: IGFW
63 REAL(8) :: IGFHTN
64
65 CGF = GAMM( IGFA )
66 IGFC1 = 1.DO + IGFP(1)*IGFA + &
67 & IGFP(2)*IGFA**2.DO + &
68 & IGFP(3)*IGFA**3.DO + &
69 & IGFP(4)*IGFA**4.DO + &
70 & IGFP(5)*(exp(-IGFP(6)*IGFA)-1.DO)
71 IGFC2 = IGFP(1) + IGFP(2)/IGFA + &
72 & IGFP(3)/IGFA**2.DO + IGFP(4)/IGFA**3.DO
73 IGFC3 = IGFR(1) + IGFR(2)*IGFA + &
74 & IGFR(3)*IGFA**2.DO + IGFR(4)*IGFA**3.DO
75 IGFC4 = IGFS(1) + IGFS(2)/IGFA + &
76 & IGFS(3)/IGFA**2.DO + IGFS(4)/IGFA**3.DO + &
77 & IGFS(5)/IGFA**4.DO
78 IGFW = 1.DO/2.DO + 1.DO/2.DO * TANH( IGFC2 * ( IGFX - IGFC3 ) )
79 IGFUNC = EXP(-IGFX) * IGFX**IGFA * &
80 & ( 1.DO/IGFA + ((IGFC1*IGFX)/(IGFA*(IGFA+1.DO))) + &
81 & (((IGFC1*IGFX)**2.DO)/(IGFA*(IGFA+1.DO)*(IGFA+2.DO))) ) * &
82 & ( 1.DO - IGFW ) + &
83 & CGF*IGFW*(1.DO-(IGFC4**(-IGFX)))
84
85 RETURN
86 END FUNCTION IGFUNC
87
88
89
90 !==== Saturation Vapor Mixing Ratios (Liquid) =====!
91 REAL(8) FUNCTION FQSATL( T, P )
92
93 REAL(8) :: T !! temperature (liquid)
94 REAL(8) :: P !! pressure
95 REAL(8), PARAMETER :: RAIR = 287.04 !! gas constant of air
96 REAL(8), PARAMETER :: RVAP = 461.DO !! gas constant of q
97 REAL(8), PARAMETER :: EMELT = 3.4D+5 !! latent heat of melt
98 REAL(8), PARAMETER :: TMELT = 273.15D0 !! melting point of water
99 REAL(8), PARAMETER :: ESO = 611.DO !! saturate e 0deg C(Pa)
100 REAL(8), PARAMETER :: EL = 2.5D+6 !! latent heat of cond.
101 REAL(8), PARAMETER :: EPSV = RAIR / RVAP !! molec. weight water/air
102
103 FQSATL = EPSV * ESO / P * EXP( EL/RVAP * ( 1.DO/TMELT - 1.DO/T ) )
104
105 RETURN
106 END FUNCTION FQSATL
107
108
109
110 !==== Saturation Vapor Mixing Ratios (Ice) =====!
111 REAL(8) FUNCTION FQSATI( T, P )
112
113 REAL(8) :: T !! temperature (ice)
114 REAL(8) :: P !! pressure
115 REAL(8), PARAMETER :: RAIR = 287.04 !! gas constant of air
116 REAL(8), PARAMETER :: RVAP = 461.DO !! gas constant of q
117 REAL(8), PARAMETER :: EMELT = 3.4D+5 !! latent heat of melt
118 REAL(8), PARAMETER :: TMELT = 273.15D0 !! melting point of water
119 REAL(8), PARAMETER :: ESO = 611.DO !! saturate e 0deg C(Pa)
120 REAL(8), PARAMETER :: EL = 2.5D+6 !! latent heat of cond.

```

```

121     REAL(8), PARAMETER :: EPSV = RAIR / RVAP !! molec. weight water/air
122
123     FQSATI = EPSV * ESO / P * EXP( (EL+EMELT)/RVAP *( 1.DO/TMELT - 1.DO/T )
124         )
125
126     RETURN
127 END FUNCTION FQSATI
128 ! ===== !

```


Acknowledgements

I would like to express my sincere gratitude to Professor Toshihiko Takemura for providing an excellent research environment, professional guidance, and encouragement throughout my research work. I also thank Professor Kazuaki Kawamoto for leading to this research field when I was a BSc student. Associate Professor Kentaroh Suzuki is one of the most important co-authors in my research works, whose insightful comments and advice have been a great help in this thesis as well. I also appreciate the feedback offered by the sub-chief examiners, Professor Hajime Okamoto, and Professor Masahiro Watanabe.

I would like to thank all members of Professor Toshihiko Takemura's laboratory for fantastic days and encouragement. Assistant Professor Nawo Eguchi gave me constructive comments and warm encouragement. I am particularly grateful for the assistance given by Ms. Risa Yoshimura, the technical staff of the Takemura laboratory. I also thank Ms. Harumi Fujii, Ms. Kyoko Kakuchi, and Ms. Noriko Kashihara for their support during my laboratory life. Mr. Tomoaki Hirakawa always provided opportunities for positive discussion from mathematical perspectives.

I would like to thank Assistant Professor Yousuke Sato at Nagoya University for helpful discussion from the perspective of cloud resolving model (Chapters 4, 5, and 6). For investigating the susceptibilities of cloud and precipitation to aerosols (Chapter 4), Dr. Christopher Ryutaro Teraï at Lawrence Livermore National Laboratory, USA, provided valuable comments for analysis approach, and Dr. Edward Gryspeerdt at Imperial College London, UK, shared information on a retrieval bias in satellites. I also thank Dr. Karsten Peters at Max-Planck-Institut für Meteorologie, Germany, for his insightful discussion about model parameterization of subgrid-scale variability. For the development of new cloud-precipitation scheme (Chapters 5), Assistant Professor Youichi Kamae at University of Tsukuba, provided the package of single-column model of MIROC, and supported to configure for the CGILS experiments. I would like to express the deepest appreciation to Dr. Tomoo Ogura at National Institute for Environmental Studies for his helpful comments on cloud ice and snow microphysics, and Associate Professor Miho Sekiguchi at Tokyo University of Marine Science and Technology for her technical advices on radiation module. I received generous support from the developers of both SPRINTARS and MIROC. Simulations by MIROC-SPRINTARS were executed with the SX-9/ACE supercomputer system of the National Institute for Environmental Studies, Japan.

I would like to express my gratitude to the Japan Society for the Promotion of Science (JSPS) KAKENHI Grant-in-Aid for Research Fellows (JP15J05544), and the Program For Lead-

ing Graduate Schools “Graduate education and research training program in decision science for a sustainable society” led by Professor Tetsukazu Yahara, for their financial supports.

Last but certainly not the least, I would like to extended special thanks to my family. Without their continuous encouragements, this work would not have been possible. I dedicate this work to my parents.

