NOTES AND CORRESPONDENCE

Impacts of Sub-grid Ice Cloud Physics in a Turbulence Scheme on High Clouds and their Response to Global Warming

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Abstract

The impacts of the saturation adjustment type approach to sub-grid-scale (SGS) ice clouds in a turbulent closure scheme on the high clouds and their response to global warming were investigated based on the radiative–convective equilibrium experiments (RCEs). This was motivated by the fact that the time scale of ice condensation is several orders of magnitude longer than that for liquid water. The RCEs were conducted with uniform sea surface temperatures over the spherical domain for the Earth’s radius without rotation using an explicit cloud microphysics and a non-hydrostatic icosahedral atmospheric model. This study revealed that suppressing the phase change effect associated with the SGS ice condensation on the buoyancy of the SGS turbulence could cause approximately a 20 % reduction of the total high cloud covers and a significantly different response of high cloud amounts to global warming due to the change in static stability near high clouds, which leads to weaker vertical heat transport at a sub-grid scale there. Since the typical value of the time scale of the ice-phase cloud is much longer than that of liquid water and the ice supersaturation is in general, using the saturation adjustment type approach for SGS ice clouds could lead to an overestimation of the effect of ice condensation for the turbulent mixing and model biases in simulations with both cloud resolving and general circulation models. The present result underlines the critical nature of the treatment of SGS ice clouds in turbulence schemes which reflects a realistic ice condensation time scale not only for a better representation of high clouds in the current climate but for an improved projection of changes of high clouds due to global warming.

Keywords high cloud; high cloud in warmer climates; sub-grid turbulence; sub-grid cloud; ice-phase cloud

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1. Introduction

Clouds are critical in the Earth's radiative energy budget (Stephens et al. 2012). Since high clouds, which are chiefly composed of ice-phase hydrometeors, are effective at trapping longwave radiation (Liou 2002), the representation of high clouds is a critical issue for the representation of climatological fields of both dynamical and thermodynamical variables in the general circulation models (GCMs).

It has been recognized that the turbulent effects influence the cloud dynamics (Squires 1958; Klaassen and Clark 1985; Grabowski 1993, 2007; Grabowski and Clark 1993) and that the turbulent closure schemes affect the representation of a variety of cloud types. Noda et al. (2010) reported that the turbulent transport process affected by sub-grid-scale (SGS) cloud strongly controlled the boundary layer cloud amount. Cheng and Xu (2011) showed the strong tie among the SGS condensation, the surface sensible and latent heat fluxes, the lower tropospheric stability, and the longwave radiative cooling for the representation of low clouds. Gasparini et al. (2019) revealed the role of the radiation–turbulence interaction for the circulation inside the anvil clouds. Ohno et al. (2019) found that the high cloud cover and its response to sea surface temperature (SST) change were sensitive to the turbulent mixing length scale, which is strongly affected by the SGS condensation process. As cloud behaviors strongly modulate the climate sensitivity in GCMs (Bony 2005; Zelinka et al. 2013; Sherwood et al. 2014), the turbulent closure scheme is a critical component for the better projection of future climate.

The representation of moist process is critical for the performance of the SGS turbulence schemes. In general, the physics of SGS cloud condensation in turbulent schemes is considered based on the two types of approaches. The first one is an all-or-nothing approach where only the value of the grid-scale humidity is used to evaluate the saturation. This approach is typically used for small grid-scale simulations (e.g., Klemp and Wilhelmson 1978; Rotunno and Emanuel 1987). The other approach is to use the SGS cloud fraction (e.g., Mellor and Yamada 1982; Bretherton and Park 2009). Mellor and Yamada (1982) proposed a scheme to calculate the SGS buoyancy flux considering SGS clouds using probability distribution functions (Mellor 1977; Sommeria and Deardorff 1977) and an assumption that the physics of cloud condensation is sufficiently fast. This “fast” condensation physics assumption (Mellor and Yamada 1982) is similar to bulk water saturation adjustment schemes

in liquid water cloud microphysics parameterizations (e.g., Wilhelmson and Ogura 1972; Soong and Ogura 1973). Several families of turbulent closure parameterization schemes commonly used in both the GCM and CRM studies rely on the SGS cloud schemes (Golaz et al. 2002a, b; Nakanishi and Niino 2009; Kuwano-Yoshida et al. 2010; Āurán et al. 2018). Chaboureaud and Bechtold (2002) attempted the extension of the SGS cloud approach of Sommeria and Deardorff (1977) and Mellor (1977) for the mixed-phase clouds. Using the results of CRM simulations with the Méso-NH (Lac et al. 2018) as pseudo observations, Chaboureaud and Bechtold (2002) proposed a diagnostic scheme of SGS mixed-phase clouds, wherein the latent heat and water vapor saturation mixing ratio in the formulas of Sommeria and Deardorff (1977) and Mellor (1977) are replaced by linear combinations of those of liquid and ice water depending on the values of the grid-scale temperature. Olson et al. (2019) employed this diagnostic scheme for the representation of the moist process of the Mellor–Yamada–Nakanishi–Niino scheme (MYNN; Nakanishi and Niino 2009).

It has been recognized that the time scale of ice condensation is generally several orders of magnitude longer than that of liquid water (Khvorostyanov and Curry 2014) and that the ice supersaturation frequently occurs (Spichtinger et al. 2003). For example, the supersaturation relaxation time in crystalline clouds with a concentration of 100 per liter, which is the typical value in the upper troposphere (Heymsfield and Miloshevich 1995; Gryspeerd et al. 2018), and the mean radius of 20 μm is 30 minutes, which is several orders of magnitude larger than the time step length typically used in the CRM simulations ($\sim 1\text{--}10$ s). Based on the above, several modern cloud microphysical schemes for CRM studies adopt the explicit calculation of ice nucleation (Hong et al. 2004; Milbrandt and Yau 2005; Morrison et al. 2005; Seifert and Beheng 2006; Roh and Satoh 2014; Seiki et al. 2015a). Since the cloud microphysics scheme (Caniaux et al. 1994) employed in the study of Chaboureaud and Bechtold (2002) adopts an ice water adjustment scheme proposed by Tao et al. (1989) to remove any ice supersaturation, and the condensation physics of ice clouds in the adopted model was sufficiently fast, the methodology of Chaboureaud and Bechtold (2002) to evaluate the ice cloudiness can cause model biases. Although statistical approaches could be useful for the representation of SGS clouds, the validity of using the fast condensation physics assumption for SGS ice clouds in the turbulent mixing processes is

questionable, specifically for CRMs. Since high-cloud behaviors can be strongly affected by the SGS turbulence, the treatment of SGS ice clouds physics in the turbulence should be designed to be more consistent with realistic physical processes; it particularly affects global scale simulation with high-resolution models, now that global cloud resolving models (GCRMs) have become more popular (Satoh et al. 2019; Stevens et al. 2019).

This note reports on the large impact of the fast condensation physics assumption for ice clouds in the turbulent closure scheme on the high clouds and their response to SST change. This study is a follow-up of earlier studies of Noda et al. (2010) and Ohno et al. (2019). Noda et al. (2010) showed that an SGS cloud scheme in a turbulence scheme had a large impact on boundary layer clouds. In this study, we examine the impact of the SGS ice cloud scheme on the behavior of high clouds based on RCEs to simplify the problem. We also review the validity of the saturation adjustment type approach for ice phase, which was introduced by Noda et al. (2010) considering the consistency in the moist process between the cloud microphysics and the turbulent closure schemes. In Section 2, the model setting, and the experimental design are described. Section 3 presents how the high clouds and their response to SST change were modulated by the SGS ice cloud scheme. Our conclusions are presented in Section 4.

2. Model setup

2.1 Experimental design

The experimental settings examined in this study inherit those used by Ohno and Satoh (2018) and Ohno et al. (2019). Numerical simulations with RCE configurations with uniform SSTs over the spherical domain for the Earth's radius without rotation were conducted using the NICAM (Tomita and Satoh 2004; Satoh et al. 2008, 2014). Cloud microphysical processes were calculated using a double-moment microphysics scheme (NDW6; Seiki and Nakajima 2014; Seiki et al. 2015a), which considers six categories of hydrometeors, namely, water vapor, cloud water, cloud ice, rain, snow, and graupel. The NDW6 scheme explicitly calculates the generation and evaporation of ice cloud particle and does not use saturation adjustment schemes for ice clouds. The MstrnX scheme (Sekiguchi and Nakajima 2008) was employed for radiative transfer calculations. The bulk surface flux was calculated following an approach of Louis (1979), Uno et al. (1995), and Moon et al. (2007). The level 2 of a modified version of the MYNN scheme (Noda

et al. 2010) was used to calculate SGS turbulence in both the planetary boundary layer and the free atmosphere.

The modified MYNN scheme employs an SGS condensation scheme for both liquid and ice clouds to calculate the virtual potential temperature and SGS turbulent buoyancy flux (Noda et al. 2010), which is similar to the MYNN scheme used in the WRF–ARW model (Olson et al. 2019). The relationship between the turbulent buoyancy flux and SGS clouds will be presented in the next subsection. To evaluate impacts of SGS ice condensation process on high clouds in detail, we examined configurations with and without an SGS condensation scheme for ice water condensate (hereafter referred to as ICE and NOICE, respectively). With the NOICE configuration, only the liquid SGS clouds are considered even with the temperature below the melting point. Since the typical value of the time scale of phase relaxation associated with ice hydrometeors is much larger than that of the time step length used in the CRM simulations, the NOICE configuration is more plausible than the ICE one from the view point of the phase change effect on the buoyancy of the SGS turbulence. In the NICAM's physics package, the SGS condensation schemes are used only for the diagnosis of the turbulent diffusivity and do not directly affect the cloud microphysics and radiative processes.

We used a 28 km horizontal grid spacing, which is the same as that used in the study of Ohno et al. (2019). The horizontal grid spacing is coarser than those used in typical CRM simulations. Previous studies using horizontal grid spacings ranging from 28 km to a sub-kilometer (Tomita et al. 2005; Satoh and Matsuda 2009; Sato et al. 2009; Iga et al. 2011; Miyamoto et al. 2013; Noda et al. 2014; Kajikawa et al. 2016; Ohno and Satoh 2018; Ohno et al. 2019; Hohenegger et al. 2020), however, demonstrated a qualitative similarity of results. The results obtained from 28-km horizontal grid simulations, therefore, can be used to investigate sensitivity to cloud processes on high cloud properties, and provide useful insights for the impact of the SGS ice condensation scheme on high cloud behavior. We used a 78 vertical layer configuration, which is similar to those used in the studies by Seiki et al. (2015b) and Ohno and Satoh (2015, 2018). The vertical layer depth of the altitude from the lower troposphere to the lower stratosphere was 400 m with this configuration.

We used a tropical climatological profile of ozone used in the studies of Ohno and Satoh (2018), Wing et al. (2018), and Ohno et al. (2019). The values described in Table 1 were employed for the concen-

Table 1. Summary of the absorption gas concentrations used in the present study.

Species of absorption gas	Concentration
carbon dioxide	348 ppmv
methane	1650 ppbv
nitrous oxide	306 ppbv
chlorofluorocarbon	100 pptv

trations of other absorption gases. A constant value of 434 W m^{-2} was used for the incoming solar insolation for the entire domain with a zero-zenith angle without a diurnal cycle. This value corresponds to the daily and annual mean of the solar irradiance at the equator. The rotation rate was set to 0. Fixed SSTs of 300 K and 304 K were employed as the bottom boundary condition.

For the initialization, we used snapshot datasets of simulations with a relatively low vertical resolution configuration. We first conducted 100-day RCE simulations with the 38-layer setting used in the study by Kodama et al. (2015). These were initialized with a zonally averaged profile at the equator obtained from the National Centers for Environmental Prediction global analysis data, corresponding to 00:00 UTC on 1 June 2004, with prescribed noises for vertical wind field in the lower 3-km layer with SSTs of 300 K and 304 K. The snapshot datasets of the simulations with SSTs of 300 K and 304 K after the 100-day time integration were vertically interpolated/extrapolated as the initial conditions of the 78-layer simulations with SSTs of 300 K and 304 K, respectively.

All simulations with the 78-layer configuration were run for 60 days. Since the simulations were initialized with snapshot datasets in the quasi-equilibrium states of the simulations with the lower vertical resolution, statistical equilibrium was reached after approximately 10 days, as in Fig. 1a. Figure 1a shows the time evolution of the globally averaged 1 day running mean precipitation rate. These are similar to those shown in the previous studies by Ohno and Satoh (2018) and Ohno et al. (2019). The energy balances at the quasi-equilibrium states in the simulations are shown in Fig. S1. Figure 1b shows the hourly averaged outgoing longwave radiation distributions at the end of day 60 for the NOICE simulations with an SST of 304 K. It can be seen that the convection organized into clusters, which is consistent with the results of earlier RCE simulations (e.g., Arnold and Randall 2015; Hohenegger and Stevens 2016; Wing et al. 2018). Thus, analyses were made over the last

50 days of the simulations.

2.2 SGS clouds in the turbulence scheme

In this subsection, we briefly review how SGS clouds are incorporated into the MYNN scheme. Here, we only consider the liquid-phase clouds for simplicity. To benefit from the conservative property under the phase change, the MYNN scheme employs the total water content q_w and the liquid water potential temperature θ_l (Betts 1973) as the prognostic variables. q_w and θ_l are defined as

$$q_w \equiv q_v + q_l, \quad (1)$$

$$\theta_l \equiv \theta - \frac{\theta}{T} \frac{L}{C_p} q_l, \quad (2)$$

where q_v is the specific humidity, q_l is the liquid water, θ is the potential temperature, T is the temperature, L is the latent heat of evaporation, and C_p is the specific heat of dry air at constant pressure. The thermodynamical fields are related to the dynamical fields through the virtual potential temperature θ_v , defined as:

$$\theta_v \equiv [1 + (\varepsilon^{-1} - 1)q_w - \varepsilon^{-1}q_l] \left(\theta_l + \frac{\theta}{T} \frac{L}{C_p} q_l \right), \quad (3)$$

$$\varepsilon \equiv \frac{R_d}{R_v}, \quad (4)$$

where R_v and R_d are the gas constants of water vapor and dry air, respectively. Extracting the fluctuating part from the above equation, the following relation among the covariances is obtained:

$$\langle \phi \theta_v \rangle = \beta_r \langle \phi \theta_l \rangle + \beta_w \langle \phi q_w \rangle + \beta_l \langle \phi q_l \rangle, \quad (5)$$

$$\beta_r \equiv 1 + (\varepsilon^{-1} - 1) \bar{q}_w - \varepsilon^{-1} \bar{q}_l, \quad (6)$$

$$\beta_w \equiv (\varepsilon^{-1} - 1) \left(\bar{\theta}_l + \frac{\theta}{T} \frac{L}{C_p} \bar{q}_l \right), \quad (7)$$

$$\beta_l \equiv [1 + (\varepsilon^{-1} - 1) \bar{q}_w - 2\varepsilon^{-1} \bar{q}_l] \frac{\theta}{T} \frac{L}{C_p} - \varepsilon^{-1} \bar{\theta}_l, \quad (8)$$

where ϕ is an arbitrary variable, and angle brackets $\langle \rangle$ and overbars $\bar{}$ represent ensemble means of turbulent and thermodynamical variables, respectively. β_l consists of both the effects of the latent heat release and water loading associated with the fluctuation of q_l , but the latter's contribution is, in general, smaller than the former.

To represent $\langle \phi \theta_v \rangle$ as a function of $\langle \phi \theta_l \rangle$ and $\langle \phi q_w \rangle$,

Fig. 1. (a) Time evolution of the globally averaged 1 day running mean precipitation rate in the simulations. The black and red lines indicate simulations with SSTs of 300 K and 304 K, respectively. The solid and dash lines indicate the ICE and NOICE simulations, respectively. (b) Hourly averaged OLR distributions at the end of day 60 for the NOICE simulations with an SST of 304 K.

Mellor and Yamada (1982) assumed a binormal distribution G for θ_l and q_w , and the physics of condensation is sufficiently fast according to $q_l = (q_w - q_s)H(q_w - q_s)$, where H is the Heaviside operator, and q_s is the saturation specific humidity. Thus, applying the assumption allows neither the state of the supersaturation nor the existence of clouds in the sub-saturation. Calculating the moments of G , the SGS cloud fraction R , the mean liquid water $\bar{q}_l$, and the covariance $\langle \phi q_l \rangle$ are expressed as:

$$R = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} H(q_w - q_s) G(\theta_l, q_w) d\theta_l dq_w, \quad (9)$$

$$\bar{q}_l = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} (q_w - q_s) H(q_w - q_s) G(\theta_l, q_w) d\theta_l dq_w, \quad (10)$$

$$\langle \phi q_l \rangle = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \phi(q_w - q_s) H(q_w - q_s) \cdot G(\theta_l, q_w) d\theta_l dq_w. \quad (11)$$

The integrals yield

$$R = \frac{1}{2} \left\{ 1 + \operatorname{erf} \left[\frac{a(q_w - q_{sl})}{2^{3/2} \sigma_s} \right] \right\}, \quad (12)$$

$$\bar{q}_l = aR(q_w - q_{sl}) + \frac{2\sigma_s}{(2\pi)^{1/2}} \exp \left[-\frac{a^2(q_w - q_{sl})^2}{8\sigma_s^2} \right], \quad (13)$$

$$\frac{\langle \phi q_l \rangle}{a\langle \phi q_w \rangle - b\langle \phi \theta_l \rangle} \equiv R' = R - \frac{\bar{q}_l}{2\sigma_s} \frac{1}{(2\pi)^{1/2}} \exp \left[-\frac{a^2(q_w - q_{sl})^2}{8\sigma_s^2} \right], \quad (14)$$

where

$$a \equiv \left(1 + q_{sl,T} \frac{L}{C_p} \right)^{-1}, \quad (15)$$

$$b \equiv a \frac{T}{\theta} q_{sl,T}, \quad (16)$$

$$\sigma_s \equiv \frac{1}{4} (a^2 \langle q_w^2 \rangle - 2ab \langle q_w \theta_l \rangle + b^2 \langle \theta_l^2 \rangle), \quad (17)$$

Fig. 2. Globally averaged 6 h average cloud cover for a) total, b) thin, c) medium, and d) thick high clouds for the ICE and NOICE simulations using SSTs of 300 (black) K and 304 (red) K. Note that the ranges of the vertical axes differ in each panel. High clouds were defined by the ISCCP cloud-type definitions of Rossow and Schiffer (1999). e) Cloud cover response to increasing SST. The purple, green, blue, and orange indicate the response for the total, thin, medium, and thick high clouds, respectively.

and q_{sl} and $q_{sl,T}$ are the specific humidity and its temperature derivative with the temperature value of T_l ($\equiv \theta_l T / \theta$). Substituting Eq. (14) into Eq. (5), we obtain

$$\langle \phi \theta_v \rangle = \beta_\theta \langle \phi \theta_l \rangle + \beta_q \langle \phi q_w \rangle, \quad (18)$$

$$\beta_\theta \equiv \beta_T - \beta_l R' b, \quad (19)$$

$$\beta_q \equiv \beta_w + \beta_l R' a. \quad (20)$$

In the case with $\phi = w$, Eq. (18) represents the turbulent buoyancy flux and the second terms in Eqs. (19) and (20) can be interpreted as the effect of the latent heat release and the water loading associated with the SGS condensation, respectively.

With the SGS ice clouds, we added the cloud ice q_i to q_w , used the ice–liquid water potential temperature θ_{il} defined as Eq. (28) of Tripoli and Cotton (1981) instead of θ_l defined as Eq. (2), and modified L and the formulas of q_s depending on the values of temperature, which is similar to Chaboureaud and Bechtold (2002). For the evaluation of L and q_s , not the value of actual

temperature T but the value of T_l was used following the manner of Mellor and Yamada (1982). The T_l is always smaller than T , but the difference between T_l and T is generally less than 2 % (Sommeria and Deardorff 1977).

3. Results

3.1 High cloud cover response

We begin by examining the cloud cover. Figure 2a shows the globally averaged high cloud cover for all simulations. High clouds were defined using the International Satellite Cloud Climatology Project (ISCCP) cloud-type definitions (Rossow and Schiffer 1999); those whose optical depths are larger than 0.3, and tops locate in altitudes between 50 hPa and 440 hPa. The high-cloud cover decreased with the SST increase for the ICE simulations. This decrease of high clouds was similar to those reported in previous studies using conventional GCMs (e.g., Zelinka and Hartmann 2010; Bony et al. 2016) but was contrary to the results using finer vertical simulations by Ohno et al. (2019) due to the vertical resolution dependency of the tur-

bulent diffusivity. Next, the cloud cover responses for each type were examined to clarify the contributions of different cloud types. We defined thin, medium, and thick high clouds as cirrus, cirrostratus, and deep convection clouds using the ISCCP cloud-type definitions. The globally averaged thin, medium, and thick cloud covers are shown in Figs. 2b–d, respectively. The high cloud decrease associated with the increase of the SST in the ICE simulations were caused by the decrease of the thin and medium clouds (Fig. 2e). The contributions of the changes of thick clouds were almost negligible. These results were consistent with those of the simulations with the 78-vertical layer configuration studied by Ohno et al. (2019).

The total high-cloud covers in NOICE simulations were approximately 20 % smaller than those in ICE simulations for both 300 K and 304 K, as shown in Fig. 2a. The difference in high-cloud cover between ICE and NOICE simulations was caused by the difference in optically thin clouds (Figs. 2b–d). Focusing on the response of the SST change, the high clouds increased in the NOICE simulations with the SST increase due to the thin-cloud increase in contrast to the decrease in the ICE simulations (Fig. 2e).

The reduction of high clouds and alternation of the sign of cloud cover response associated with suppressing the SGS ice condensation in the turbulent closure scheme resemble the impacts of increasing vertical resolutions reported in the study of Ohno et al. (2019). They determined that the vertical resolution dependency of high clouds and their response to SST change were related to the turbulent mixing near high clouds. The results suggest that the application of the SGS ice condensation in the turbulent scheme changed the high clouds in the quasi-equilibrium states by altering turbulent mixing near high clouds.

3.2 Turbulent diffusivity

To clarify the impacts of the SGS ice condensation on the turbulent mixing, we investigate the turbulent diffusivity fields. Figure 3a shows binned vertical profiles of the turbulent diffusivity calculated with the SGS ice condensation sorted by the ice water path (IWP). The turbulent diffusivity and IWP were calculated using a snapshot dataset of the ICE simulation with a 300 K SST at the end of the integration time. The IWP was defined as the vertically integrated mass concentrations of cloud ice and snow. Since the graupel particles tend to have large mass concentrations and small optical effects, the inclusion of the graupel for the IWP calculation can blur the correspondence between the IWP and the cloud optical thickness.

Thus, the graupel was excluded for the calculation of the IWP. The horizontal axis shows the area percentile of the IWP bins; 100 % corresponds to the core of deep convective regions, and smaller values correspond to clearer regions with a smaller IWP.

The regions with large values of diffusivity can be seen not only inside the convective core region where the IWP values exceeded the value of the approximately 97th percentile but also just outside of the convective core region. The peak height of the turbulent diffusivity was near a 12.5 km height, which is just above the peak height of the ice water condensate (~ 12 km). Although the values of turbulent diffusivity near the convective core tend to be larger than those reported by Ohno et al. (2019) using a 4 h averaged dataset, the spatial relationship between the turbulent diffusivity and the ice water condensate was consistent with Ohno et al. (2019).

Figure 3b is similar to Fig. 3a but without the SGS ice condensation for the evaluation of the diffusivity. Since level 2 of the MYNN scheme diagnoses the turbulent diffusivity only from the grid-scale variables, the impacts of the SGS ice condensation on the turbulent diffusivity promptly emerges. Although maxima were inside and just outside the convective core region, as the case with the SGS ice condensation, the peak values without the SGS ice condensation were considerably smaller than those with the SGS condensation. The peak altitudes shifted upward by suppressing the SGS condensation. Similar differences can be seen in the diffusivity profiles with a 304 K SST, as shown in Figs. 3c and 3d.

To understand why we see a large difference, the relationship between the turbulent diffusivity and the SGS ice condensation scheme was considered. In the Mellor–Yamada moist turbulent closure scheme, the turbulent diffusivity K is evaluated from the product of the three variables: the stability function S , the mixing length scale L , and the square root of the doubled turbulent kinetic energy (TKE) q ($= (2\text{TKE})^{1/2}$), or the turbulent velocity scale.

$$K = LqS. \quad (21)$$

The S is a function of the gradient Richardson number Ri , which is a ratio of the square of the amplitude of the vertical wind shear and the square of the Brunt–Väisälä frequency N including the effect of the SGS ice condensation. The N was defined as:

$$N \equiv \left[\frac{g}{\theta} \left(\beta_\theta \frac{\partial \theta_l}{\partial z} + \beta_q \frac{\partial q_w}{\partial z} \right) \right]^{1/2} \left(= \left(\frac{g}{\theta} \frac{\partial \theta_v}{\partial z} \right)^{1/2} \right), \quad (22)$$

Fig. 3. Binned vertical profiles of the turbulent diffusivity (color) calculated a) with and b) without the SGS ice condensation scheme, respectively, sorted by the IWP. The dynamical and thermodynamical fields used were those of a snapshot of the ICE simulation with a 300 K SST at the end of the integration time. The white and black lines indicate the turbulent diffusivity and the ice condensate, respectively. The contour intervals for the frequency and ice condensate are 2 % and $20 \text{ m}^2 \text{ s}^{-1}$, respectively. c) and d) are the same as a) and b) for the simulation with a 304 K SST.

where g is the gravitational acceleration, z is the height, β_θ and β_q are the sums of the correction terms relevant to the sub-grid condensation and the differentiation of θ_v regarding θ_i and q_w defined as Eqs. (19) and (20), respectively (Mellor and Yamada 1982). In their level 2 scheme, the TKE is diagnosed from the balance among the buoyancy production, shear production, and dissipation. The dissipation term is proportional to L^{-1} . The L is calculated as a harmonic average of three length scales as:

$$\frac{1}{L} = \frac{1}{L_S} + \frac{1}{L_T} + \frac{1}{L_B}, \quad (23)$$

where L_S is the length scale in the surface layer, L_T is a length scale of the atmospheric boundary layer, and

L_B is the buoyant length scale. In the free atmosphere, L is, in general, dominated by L_B (Ohno et al. 2019), which is proportional to the inverse of N . Since S , q , and L were controlled N , the static stability including the effect of the SGS ice condensation, or the value of the N , is critical to determine K .

To examine the impact of the SGS ice condensation scheme on static stability, the frequency of the occurrence of static instability was investigated. Figures 4a and 4b present binned vertical profiles of the frequency of the occurrence of static instability ($N^2 < 0$) for the simulation with an SST of 300 K calculated with and without the SGS ice condensation scheme. Figures 4a and 4b showed that the static instability frequently occurred near the convective region in both cases with and without the SGS ice condensation

Fig. 4. Binned vertical profiles of the frequency of occurrence of static instability (color) calculated a) with and b) without the SGS ice condensation scheme, respectively, sorted by the IWP. The dynamical and thermodynamical fields used were those of a snapshot of the simulation using the partial condensation for ice clouds with a 300 K SST at the end of the integration time. The white and black lines indicate the frequency and the ice condensate, respectively. The contour intervals for the frequency and ice condensate are 2 % and 5 mg kg^{-1} , respectively. c) and d) are the same as a) and b) for the simulation with a 304 K SST.

scheme. It can be seen that the frequencies with the SGS ice condensation scheme were much larger than those without the scheme. The distributions of the frequency of the occurrence of static instability in Figs. 4a and 4b were consistent with those of the turbulent diffusivity shown in Figs. 3a and 3b, respectively. The similar differences between the cases with and without the SGS ice condensation scheme can be seen in the simulations with an SST of 304 K (Figs. 4c, d). These indicate that the application of the SGS ice condensation in the turbulent scheme changed the high cloud covers and their response to SST change by modulating the static stability near the convective core regions.

4. Discussion and summary

This study investigated the impacts of the “fast” condensation physics assumption for ice clouds, which was originally proposed for liquid-phase clouds (Mellor and Yamada 1982), or the inclusion of ice phase as part of the fast adjustment, in the turbulent closure scheme on high clouds and their response to SST change based on the RCEs. The sensitivity experiments revealed that the suppression of the SGS ice cloud scheme caused approximately a 20 % reduction of the total high cloud cover in the simulations with SSTs of both 300 K and 304 K and the alternation of the sign of cloud cover response to the SST change. It was also determined that the SGS ice cloud scheme strongly altered the static stability near high clouds.

The reduction of the static stability and/or the occurrence of static instability can enhance the TKE production and elongate the turbulent mixing length, resulting in the enlargement of the turbulent diffusivity. The enlargement of the turbulent diffusivity was seen in the case with the SGS ice cloud scheme. The comparison of the distributions of the turbulent diffusivity and the occurrence of static instability showed considerable correspondence. These indicate that the application of the SGS ice condensation in the turbulent scheme changed the high cloud covers and their response to SST change by modulating the static stability near high clouds, which was similar to the impacts of increasing vertical resolution, as reported by Ohno et al. (2019).

Although the time scale of the ice condensation physics depends on both number concentration and the size of ice crystal, the time scale of ice condensation is, in general, several orders of magnitude longer than that of liquid water (Khvorostyanov and Curry 2014) and the ice supersaturation is a frequent phenomenon (Spichtinger et al. 2003). Thus, SGS ice cloud schemes similar to the saturation adjustment approach overestimate the SGS ice cloud fraction, particularly when simulations are conducted using cloud schemes that allow the ice supersaturation condition. Although SGS ice clouds were not used for the radiative transfer calculations in the model used in this study, the radiative effects of SGS ice clouds should be overestimated, if the saturation adjustment type SGS ice cloud schemes are coupled with radiative transfer schemes. Furthermore, as the impact of the phase change of ice-phase clouds on the dynamical fields appears more gently than those of the liquid-phase clouds, applying such SGS ice cloud schemes for SGS turbulent schemes exaggerates the SGS buoyancy flux in the regions with temperature below freezing point. These indicate that using the saturation adjustment type SGS ice cloud schemes could cause model biases in simulations with not only CRMs but also GCMs. Furthermore, since the typical value of the time scale of phase relaxation associated with ice hydrometeors is much larger than that of the time step length used in the CRM simulations, the validity of applying statistical approaches is questionable from the stationarity states viewpoint in principle. The performance of the turbulent closure scheme without the SGS ice cloud scheme at the altitude of high clouds, therefore, should be more plausible as a single component than that with the scheme in principle, although a realistic performance should exist between the two.

Recently, a number of GCRMs have been emerging

(Stevens et al. 2019; Satoh et al. 2019). Additionally, aided by the improvement of computational power, climatological studies employing models with explicit cloud physics have been increasing (Stan et al. 2010; Wyant et al. 2012; Kodama et al. 2015; Haarsma et al. 2016; Noda et al. 2019). Since the representation of high clouds is critical for the representation of climatological fields, and high cloud behaviors can be strongly affected by the SGS turbulence, the improvement of the treatment of SGS ice clouds physics in the turbulence should be desirable.

Supplements

Figure S1 shows a globally averaged a) the latent heat, b) sensible heat, c) longwave, and d) shortwave fluxes at the surface and e) the longwave and f) shortwave fluxes at the top of the atmosphere, respectively, at the quasi-equilibrium states.

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