Evaluation of the Influence of Saturation Adjustment with Respect to Ice on Meso-scale Model Simulations for the Case of 22 June, 2002

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Abstract
A meso-scale model simulation is sometimes conducted with a saturation adjustment scheme with respect to ice in the upper troposphere. In such a simulation, the super-saturation with respect to ice is not permitted in upper tropospheric clouds, which is contradicting with natural situation.

In this study, we evaluate the influence of the application of ice-saturation adjustment to upper tropospheric clouds in a meso-scale model. When the ice-saturation adjustment is applied, all of the excess water vapor in upper troposphere is consumed to produce large amount of cloud ice and subsequently snow. Small amount of excess water vapor remain for the depositional growth of snow. Since the fall velocity of snow becomes smaller, the residence time of snow in the atmosphere becomes longer. Consequently, the surface rainfall intensity, on domain-average, becomes slightly smaller through the integration time, compared to the simulation without ice-saturation adjustment.

1. Introduction

It is generally assumed in meso-scale model simulations that water vapor in the atmosphere does not exceed the saturation with respect to liquid water. This is a good approximation, because the super-saturation with respect to liquid water, which can be realized in natural clouds, is at most a few percents due to quick consumption of excess water vapor by a large number of cloud condensation nuclei (CCN). This is the basis for the application of water-saturation adjustment scheme to meso-scale model simulations. On the other hand, the super-saturation with respect to ice often becomes ten percents or more in natural situations. Heymsfield et al. (1998) reported the ice-super-saturation of several tens percents in the upper troposphere. Those fundamentally come from that the number of ice nuclei (IN) is much smaller than that of CCN in the atmosphere.

Some meso-scale models have an ice-saturation adjustment scheme to be applied to the upper troposphere. With the ice-saturation adjustment, all of excess water vapor is converted into cloud ice in upper tropospheric clouds during one time step in model simulations. This is supposed to lead the overestimation of solid water condensates in the atmosphere. Khvorostyanov and Sassen (1998) evaluated the super-saturation with respect to ice which can be realized in cirrus clouds using their 2D model with spectral (bin) microphysics. They concluded that the super-saturation with respect to ice long remained in the magnitude of 10 %.

The NonHydrostatic Model (NHM) of the Japan Meteorological Agency (JMA), presented in Saito et al. (2006), is in use for the operational forecast with the option of ice-saturation adjustment scheme for the temperature colder than –36°C (Yamada 2003). The Weather Research and Forecasting Model (WRF) also has a corresponding scheme (Skamarock et al. 2005) which is basically the same as NHM for this temperature range. The cloud-resolving regional climate model developed on the basis of NHM was employed to address the change in the characteristics of Baiu front over East Asia due to global warming (Yoshizaki et al. 2005). This model was also run with that scheme. The purpose of the present study is to evaluate how the ice-saturation adjustment influences the results of numerical simulation with meso-scale models.

2. Numerical model

The meso-scale model used in the present study is based on the NHM of JMA. For cloud and precipitation processes, the NHM explicitly calculates the microphysical processes of five categories of liquid and solid water substances: cloud water, rain, cloud ice, snow, and graupel. Cumulus parameterization is not used in this study, although it is applied to the JMA operational forecast with 5-km horizontal resolution. The two-moment bulk parameterization scheme, which prognoses both the mixing ratio and number concentration, is applied to the five categories to simulate the interactions among the hydrometeors. The gravitational sedimentation of cloud ice is included. Homogeneous freezing nucleation is assumed to be active for the temperature colder than –36°C. Ice formation through deposition nucleation is implemented following Meyers et al. (1992), in which nucleation number is equated in exponential form of ice-saturation ratio. In addition, we introduced the thermodynamical effects on the ice-saturation ratio that comes from the adiabatic cooling and the water vapor consumption due to nucleation and diffusional growth of hydrometeors in an ascending air parcel (Pruppacher and Klett 1996), but this is not essential in the present study.

In the saturation adjustment scheme implemented in NHM, the water vapor surplus to ice-saturation (water-saturation) is converted to cloud ice (cloud water) for the temperature colder (warmer) than –36°C, although it should be natural procedure that the upper limit of water vapor is set to water-saturation even in the temperature colder than –36°C. Our method is partly different from that of WRF based on Tao et al. (1989), where the saturation vapor mixing ratio is defined as a mass weighted combination of saturation values over liquid water and ice between the temperatures 0°C and –40°C. However, things should essentially go the same as the other scheme as far as the temperature is colder than –40°C.
3. Design of the numerical experiment

The model domain covers the area of 300 km × 300 km centered at Kyushu Island, located in western part of Japan, with the horizontal resolution of 5 km. The top height of the model domain is about 22 km. The number of variable vertical layers is 48 in the simulations. Time integration up to 24 hours is conducted with a time step of 12 s from the initial time of 21 JST, 21 June, 2002 to simulate the Baiu precipitation system, for which in-situ measurements with aircraft were done. Since the Baiu frontal system is largely associated with ice-phase processes, it is worthwhile to take this case for a comparison of model schemes. The NHM was run in the one-way nesting manner, with the initial and hourly boundary data provided from the simulation with the 20-km mesh Regional Spectral Model (RSM) of JMA.

For control experiment, the simulation was conducted without the ice-saturation adjustment scheme (NoSA). As a sensitivity experiment, another simulation was conducted with the ice-saturation adjustment scheme in the range of temperature colder than −36°C (SA).

4. Results

4.1 Differences in microphysical characteristics

Both simulations of NoSA and SA roughly reproduced the distribution of precipitation extending several hundreds km and the intensity up to 50 mm h⁻¹ from radar observation. Figure 1 shows the horizontal distribution of number concentration of ice particles (including cloud ice, snow, and graupel) at the height of 12.6 km (~−50°C), at the forecast time FT = 15 h (12 JST, 22 June, 2002). In the NoSA experiment (Fig. 1a), the number concentration is on the order of magnitude 10 L⁻¹ and ice cloud regions are extending 100 to 200 km in a meridian direction (yellow in Fig. 1a). The several embedded areas with the values of several hundreds L⁻¹ correspond to updraft cores. The simulated microphysical properties of clouds are in good agreement with the results of in-situ air-borne measurements (Murakami et al. 2005). The aircraft flew along the longitude of 130°E from 31.5 to 28.5°N (1120-1140 JST) at the height of 12.6 km. The observed horizontal scale of cloud region with the number concentration of ice particles on the order of magnitude 10 L⁻¹ (including some embedded areas of higher number concentration) was about 150 km. In the SA experiment (Fig. 1b), the number concentration of ice particles is on the order of 10¹ L⁻¹ in the region extending more than 200 km along the meridian direction (orange in Fig. 1b). It is clear that the area of high number concentration is overestimated due to the application of ice-saturation adjustment. The mass contents of ice particles predicted in those experiments are not different very much and are slightly overestimated, compared with the observation.

Figure 2 shows the time series of total ice water amount over the calculation domain. The overestimation up to about 10 % is also found through most of integration time in the SA experiment. This result is consistent with the statement of Khvorostyanov and Sassen (1998) that the ice-saturation adjustment can lead to considerable overestimation of ice water content. Figure 3 shows the vertical distributions of the mixing ratio and number concentration of cloud ice and snow averaged over the domain and time from FT = 6 to 24 h. (The amount of graupel was quite small.) In the NoSA experiment, the peaks of mixing ratio and number concentration of cloud ice (blue solid lines in Fig. 3) are located around a height of 15 km. In the SA experiment, in contrast, the vertical profile of mixing ratio of cloud ice (blue broken line in Fig. 3a) has a peak in the height of 11 km (about −36°C). The mixing ratio of snow above 10-km height is slightly larger (red broken line in Fig. 3a) in the SA experiment than in the NoSA experiment (red solid line in Fig. 3a). The number concentration of snow (Fig. 3b) shows a drastic change, an increase of more than one order of magnitude, from the NoSA (red solid line) to SA (red broken line) experiment. The changes in mixing ratio and number concentration due to the ice-saturation adjustment bring the underestimation of the representative size of snow at most of the levels above melting level in the troposphere by several
tens percents. Since the difference in the representative size of snow is reflected in their fall velocity, the application of ice-saturation adjustment scheme in upper troposphere affects the precipitation efficiency in meso-scale model simulations.

4.2 The processes inducing the overestimation of snow

Figure 4 shows the vertical distribution of the appearance probability of ice-saturation ratio. In the SA experiment (Fig. 4b), super-saturation with respect to ice is suppressed above an 11-km height, while, in the NoSA experiment (Fig. 4a), super-saturations of up to 25% are predicted. This difference brings about a critical effect on the depositional growth of snow particles. In the SA experiment, the excess water vapor with respect to ice has been all consumed by the increment of cloud-ice mixing ratio through the ice-saturation adjustment. This prevents the increase of number concentration of cloud ice through deposition nucleation, because the super-saturation with respect to ice has become zero. Eventually, the representative size of cloud ice increases fast and reaches its prescribed upper limit (120 μm in diameter) easily. That results in large production of snow through the auto-conversion from cloud ice to snow that is dependent on the representative size of cloud ice, as shown in Fig. 5 (red broken line). The produced snow particles, however, can not increase their size through depositional growth above 11-km height, because super-saturation with respect to ice does not exist any more. This situation keeps the representative size of snow to be small (around 100 μm). In the NoSA experiment, the depositional growth of cloud ice is explicitly computed, the representative size of cloud ice grows moderately and the conversion of cloud ice to snow is slow (red solid line in Fig. 5). The produced snow particles increase their size faster through depositional growth above 11-km height (blue solid line in Fig. 5). In natural clouds, the excess water vapor with respect to ice should be used not only for the production and growth of ice crystals but also for the depositional growth of snow, simultaneously. This shortcoming prevents the snow particles from growing larger in upper troposphere, and makes the fall velocity slower. Consequently the residence time of snow particles in the atmosphere becomes longer. The abovementioned processes result into further overestimation of mixing ratio and number concentration of snow.

4.3 Influences on precipitation

Figure 6 shows the domain-averaged downward mass flux of snow at a height of 6 km (red) and 3-hour precipitation at the surface (blue). It is clear that the downward mass flux of snow is several percents smaller for the SA experiment through the simulation (red lines in Fig. 6). This feature comes from that gravitational sedimentation is less effective in the SA experiment so that the residence time of snow becomes longer. This result indicates that when the ice-saturation adjustment is applied to the simulation of winter storm case, the surface snowfall intensity could be weakened by several percents. The same tendency is also found in the rainfall intensity at the surface (blue lines in Fig. 6), although the influence is much limited. The surface rainfall intensity for the SA experiment is slightly smaller through the simulation. The application of ice-saturation adjustment gives finite influence on the surface rainfall intensity.
4.4 Dependence on horizontal resolution

Additional simulations with the horizontal resolution of 2 km were also conducted in order to examine the resolution dependence. It was found that the horizontal scale of spatial distribution of number concentration (~10 L⁻¹) of ice particles observed by aircraft was well reproduced, regardless of resolution 5 or 2 km in the NoSA experiment, although the higher resolution was in better agreement with the observation. Similar results were obtained with both resolutions (not shown) in the microphysical characteristics corresponding to Figs. 3 and 5.

5. Summary

The present study focused on the ice-saturation adjustment, in order to evaluate its influence on simulation results. Without the ice-saturation adjustment, the simulation results are in good agreement with in-situ airborne microphysical measurements. With the ice-saturation adjustment, the number concentration of ice particles and horizontal extent are overestimated very much.

According to the results, some implications may be available for the influence of ice-saturation adjustment on the actual simulations. When the purpose of simulation is to describe the inner structures of clouds and to compare them with observations, the amount and horizontal extent of snow should be regarded to be overestimated, especially for the number concentration. If one concerns the water budget through the formation of upper tropospheric clouds, special care is needed, since the humidity in upper troposphere is affected very much in the simulation. The influence on the surface precipitation may become much more obvious in the simulation of winter time precipitation systems associated with deep disturbances.

We applied a two-moment parameterization scheme to the prediction of hydrometeor. It is possible that the effect of ice-saturation adjustment becomes different from that in the present study, if the one-moment scheme is adopted as in the WRF, instead of the two-moment scheme. This is because the expressions of the growth of cloud ice and snow and the conversion rate from cloud ice to snow are quite different between both schemes. However, the simulation result should be affected through the budget of water vapor by application of ice-saturation adjustment.

The investigation of the ice-saturation adjustment should be extended to the simulations of precipitation systems in other seasons in order to statistically evaluate its adverse effects in various cloud systems. In addition, the influences of ice-saturation adjustment on the cloud radiation processes also should be addressed in the future.

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