Impact of aerosol and freezing level on orographic clouds: A sensitivity study

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Abstract

The response of clouds and precipitation to changes in aerosol properties is variable with the ambient meteorological conditions, which is important for the distribution of water resources, especially in mountain regions. In this study, a detailed bin microphysics scheme is coupled into the Weather Research and Forecasting (WRF) model to investigate how orographic clouds and precipitation respond to changes in aerosols under different thermodynamic profiles. The model results suggest that when the initial aerosol number concentration changes from a clean continental background (4679 cm^{-3}) to a polluted urban environment (23600 cm^{-3}), the accumulated surface precipitation amount can be increased up to 14% mainly due to the enhanced riming process which results from more droplets of 10-30 μm in diameter. When the freezing level is lowered from 2.85 km to 0.9 km (above 1000 hPa level), the growth of ice-phase particles via riming process is enhanced, leading to more precipitation.

However, the response of surface precipitation amount to increase in aerosol particle concentration is not linear with lowering freezing level, and there is a maximum precipitation enhancement caused by aerosols (about 14%) as the freezing level is at 1.4 km. Further sensitivity tests show that, the response of riming growth to increase in aerosol particle concentration becomes more significant with lowering the freezing level. But this effect becomes less significant as the freezing level is further lowered due to the limited liquid water. Moreover, the growth of raindrops through collision and coalescence is suppressed with lowering freezing level, due to the shorter distance between the melting level and the ground.

1. Introduction

Anthropogenic aerosols have increased substantially along with the development of industry. Atmospheric aerosols can influence the radiative budget of the climate system by absorption and scattering of solar and terrestrial radiation. Also, they can act as cloud condensation nuclei (CCN) or ice nuclei (IN) to affect cloud albedo and cloud microphysics structure, thus modifying precipitation. It is generally known that the effect of aerosols on precipitation depends on the conditions of the environment and cloud types (Khain et al., 2005; Tao et al., 2007; Khain, 2009; Cui et al., 2011).

It is important to understand precipitation modification via aerosols due to their impacts on water resources, particularly in mountains regions. These regions often provide water availability through water storage in snowpack (Kim et al., 2015). Orographic precipitation is one of the major forms of precipitation over the world (Houze, 2012). Although its response to changes in atmospheric aerosols has been studied extensively in field observations and modeling in recent years (e.g. Borys et al., 2003; Givati and Rosenfeld, 2004; Lynn et al., 2007; Alpert et al., 2008; Muhlbauer and Lohmann, 2008; Halfon et al., 2009; Saleeby et al., 2013; Fan et al., 2014; Xiao et al., 2015), there are still disputes due to the complicated dynamical and microphysical processes. Carrió and Cotton (2014) showed that increasing CCN concentrations leads to an enhancement in orographic precipitation in a cloud with lower cloud base. Muhlbauer and Lohmann (2008) found that in warm-phase

60 orographic clouds an increase in the concentration of aerosol particles leads to a
61 decrease in precipitation on the windward side of the mountain and an enhancement
62 on the leeward side, and the changes in precipitation also depend on the geometry of
63 the mountain and the dynamical flow regime, such as flow blocking, which describes
64 a situation that oncoming airflow cannot easily rise over the terrain and may become
65 blocked or dammed. In this case, the airflow may turn to a flow parallel to the barrier
66 or around an isolated mountain (Houze 2012). In addition, relative humidity and
67 vertical velocity also play a role in the aerosol effect on orographic precipitation
68 (Lynn et al., 2007; Khain, 2009).

69 As anthropogenic and natural aerosols enter into clouds and interact with them,
70 the size distribution of cloud droplets will be influenced and redistributed, thus affect
71 cloud macro- and micro-structure (Rosenfeld, 1999; Stevens and Feigold, 2009; Xue
72 et al., 2012). Observational studies suggest that, because of air pollution, orographic
73 precipitation over the windward side of mountains might be inhibited due to smaller
74 collision efficiency between drops (Givati and Rosenfeld, 2004) or accretion rate of
75 cloud water by ice-phase particles (riming process) (Borys et al., 2003). In general,
76 the spectra of cloud droplets shift toward smaller sizes as the number concentration of
77 aerosol particles increases, suppressing the formation of large drops by delaying the
78 onset of collision and coalescence (e.g. Twomey et al., 1984; Rosenfeld, 2000; Xiao et
79 al., 2014). Mitchell et al. (1990) showed that accreted cloud water (riming) comprises
80 up to about 30-40 percent of the deposited snow during moderately rimed snowfall.
81 Muhlbauer et al. (2010) suggested that the response of riming to changes in aerosols
82 varies with cases and models. They showed that the response of riming growth to
83 changes in aerosol particle concentration is different under three models (Consortium
84 for Small-Scale Modeling, Weather Research and Forecasting model, University of
85 Wisconsin modeling system), and even if the same model is used, the response of
86 riming growth to changes in aerosols is still different under different thermodynamic
87 conditions. Under the conditions with limited available vapor, a spillover effect of
88 snowfall is noticed when the aerosol particle concentration becomes higher, and is
89 believed to a large extent determined by the efficiency of riming (Saleeby et al., 2013).

90 However, small droplets formed under higher CCN concentration are more easily to
91 be transferred to the layers above the freezing level, thus increase the concentration of
92 supercooled cloud droplets and promote the ice-phase microphysical processes, such
93 as riming and the Wegener–Bergeron–Findeisen process (ice particles grow with the
94 expense of liquid droplets) (Saleeby et al., 2013; Muhlbauer et al., 2010; Xiao et al.,
95 2015). Based on a detailed bin microphysics scheme, Xiao et al. (2015) suggested that
96 the growth of ice particles through riming process became more effective in
97 orographic cloud under polluted condition, due to the increased number of droplets in
98 the size range of 10-30 μm in diameter, leading to enhanced precipitation.

99 However, the types of hydrometeor and the related microphysical processes may
100 change under different environmental conditions such as the height of the freezing
101 level changing from 500 mb to 1000 mb (Colle and Zeng, 2004). By using a
102 Numerical Weather Prediction model, Planche et al. (2014) showed that the sensitivity
103 of the surface precipitation amount to cloud droplet number concentration (CDNC) is
104 related to the cases with different freezing level. Additionally, the response of warm
105 and ice-phase microphysical processes to enhanced CCN is different and even
106 opposite (Tao et al., 2012). Muhlbauer et al. (2010) suggested that thermodynamic
107 factors play an important role in the effect of aerosol perturbations on cloud
108 microphysics, but they did not delve deeply into the aerosol effects under different
109 freezing levels. Thus, the freezing level, which can represent the proportion of
110 warm-phase and ice-phase portion in a cloud, should be considered when discussing
111 aerosol effects. Previous numerical studies of aerosol-cloud interactions mainly
112 focused on convective clouds, and very few studies have been done on orographic
113 clouds, especially by using spectral bin microphysics scheme (e.g., Fan et al, 2014).
114 Hence, the main objective of this paper is to explore the sensitivities of orographic
115 clouds to changes in aerosol particle condition and the height of freezing level. For
116 this purpose, the Weather Research and Forecasting (WRF) model with spectral bin
117 microphysics scheme is used.

118 In our previous work (e.g., Xiao et al. 2014, 2015), the contribution of ice-phase
119 microphysical processes to formation of precipitation becomes more important when

the height of the topography becomes larger. However, more studies are needed to understand the effect of aerosol particles on orographic cloud developed under different thermodynamic conditions. This study is an extension of our previous work of Xiao et al. (2015) in which the mixed-phase orographic clouds were conducted for one specific thermodynamic condition. As the freezing level is elevated or lowered, the cloud may be dominated by liquid-phase or ice-phase microphysical processes. The main goal of this study is to investigate how the effect of aerosol particles on orographic cloud and precipitation is sensitive to the changes in the thermodynamic conditions (the height of freezing level, more specifically), where the clouds are formed. To obtain more detailed information on the evolution of various cloud particles, the WRF model coupled with a detailed bin microphysics schemes is used, which can represent the natural evolution of hydrometeors without assuming of a certain size distribution function, as bulk microphysical schemes do.

2. A brief description of the model

A detailed bin microphysics scheme which has been widely used in cloud models (e.g., Reisin et al., 1996; Yin et al., 2000) coupled to the WRF model (Version 3.1) is used in this study. In the microphysics scheme, the mass and number concentrations of four hydrometeor species (liquid drops, ice crystals, snowflakes, and graupel) are included. The hydrometeor particles are described by 34 mass doubling bins covering a mass range from 0.1598×10^{-13} kg to 0.17468×10^{-3} kg (3.125 - 8063 μm in diameters for liquid drops). In addition, aerosol particles are divided into 43 bins with a radius range of 0.001-15.75 μm . In this study, aerosol particles are considered as CCN only, and our aim is to investigate the response of microphysical processes to CCN concentration. Additionally, because of the complication of aerosol composition and their mixing state, the characteristics of aerosols are simplified. The variations of the chemical composition of the aerosol particles are not considered. The CCN particles are simply assumed as a fraction of the total aerosol particle concentration and are water-soluble. In addition, all the CCN particles are assumed to be composed of ammonium sulfate (Yin et al., 2005). Hence, other components of aerosol particles,

such as organic, black carbon and nitrate, are not considered in this study. The warm-phase microphysical processes included are nucleation, condensation, evaporation, stochastic collision-coalescence, and breakup of drops. The ice-phase microphysical processes comprise drop freezing, ice nucleation (condensation freezing, deposition, immersion freezing and contact freezing), sublimation and deposition of ice-phase particles, ice multiplication, interactions of ice-drops and ice-ice, and sedimentation of both liquid drops and ice-phase particles. The Hallet-Mossop mechanism for secondary ice production is formulated base on Mossop (1978). Snow is resulted from aggregation of ice crystals in the original model (Yin et al., 2000). In order to compare the results with other previous studies, ice crystals with diameter larger than 400 μm are converted to snow according to Morrison and Gettelman (2008). Additionally, ice nucleation including vapor deposition and condensation freezing is formulated by Cooper (1986). Ice nucleation by contact freezing and immersion freezing are formulated by Cotton et al. (1986) and Bigg (1953), respectively (see Yin et al., (2000) and Xiao et al., (2015) for more details).

3. Experimental design

Idealized two-dimensional (2D) simulations of flow over a bell-shaped mountain are conducted to investigate how the effect of aerosol particles on orographic clouds is sensitive to changes in the height of freezing level. To elucidate the effects of aerosol on cloud microphysics and precipitation, other physics schemes, including the parameterizations of radiation, surface processes, and boundary layer are purposely turned off. The model domain includes 800 grid points with horizontal resolution of 500 m. To better simulate the vertical structure of the orographic clouds, 62 terrain-following levels in vertical direction are constructed with vertical grid spacing varying from 35 m at the surface to about 1850 m at the top. The time step is 2 s and the total simulation time is 8 hours. The idealized bell-shaped mountain is located horizontally at the center of the model domain with peak height of 1500 m and half-width of 20 km (Eq.(1) in Xiao et al., 2015). The initial soundings of temperature

and dew-point temperature used in the simulation are shown in Fig. 1. The horizontal wind is set to a constant value of 15 m s^{-1} below 10 km and is increased upward linearly with the rate of $1.84 \text{ m s}^{-1} \text{ km}^{-1}$. The temperature increases with the logarithm of pressure ($\partial T/\partial \ln P = 42 \text{ K}$) and the dry Brunt-Väisälä frequency is constant ($N_d = 0.0115 \text{ s}^{-1}$) with height, according to Muhlbauer and Lohmann (2008). An orographic cloud forms when the airflow approaches the mountain and is forced to rise at the windward side in a stable environment (Houze, 2012). In order to explore the sensitivity tests with different heights of freezing level, the temperatures at 1000 hPa ($T_{00\text{hPa}}$) are set to 279 K, 282 K, 285 K, 288 K, and 291 K, respectively, corresponding to the height of freezing level (FL) of 0.9 km, 1.4 km, 1.9 km, 2.35 km, and 2.85 km (above 1000 hPa, similar hereafter).

The profile of the relative humidity is represented as

$$RH(z) = c_1 + \frac{c_2 - c_1}{1 + \exp[-c_3(z - z_0)]}, \quad (1)$$

with $c_1=0.9$ (relative humidity at 1000 hPa), $c_2=0.03$, $c_3=0.0015 \text{ m}^{-1}$ and $z_0=6000 \text{ m}$. The parameters in Eq.(1) are the same in all the simulated cases. Hence, the dew-point temperature is also increased with increasing temperature. The initial aerosol particle spectra based on the field measurements from Huangshan Mountain and Beijing city to represent typical clean continental conditions and polluted urban conditions with an assumption of the fraction of water-soluble material of 34% (Wu et al., 2008; Tan, 2012; Wen, 2013), is shown in Fig. 2. Additionally, the initial number concentration of aerosol particles is assumed to decrease exponentially with a scale height of 2 km above the ground level (Yin et al., 2000). The aerosol concentration changes over entire domain under different conditions and the aerosol flows into the domain from the left boundary. In addition, this work focuses on the evolution of the cloud microphysical processes at the windward side of the mountain.

To investigate aerosols effects on cloud, two groups of sensitivity tests are constructed, “the clean background” and “the polluted environment”. The clean background is calculated using the aerosol measurements from Huangshan Mountain which represents a typical clean continental condition in eastern China. In order to

investigate the effect of anthropogenic pollution on orographic precipitation, the aerosol concentration in polluted environment is based on the measurement data from Beijing city, which is about five times that from the clean continental background. The total concentrations of aerosol particles measured on the Huangshan Mountain and in Beijing city are 4679 cm^{-3} and 23600 cm^{-3} , respectively. The size range of the instruments used for aerosol measurements is $0.01\text{-}10 \text{ }\mu\text{m}$ in diameter (Wu et al., 2008; Tan et al, 2012). In addition, the temperature at 1000 hPa changes from 279 K to 291 K with a 3 K interval to represent the changes of the freezing level with constant lapse rate and constant parameters for relative humidity profile. Hence, the dew-point temperature is also changed with temperature (Fig. 1). The parameters in the experiments are listed in Table 1.

4. Results and discussion

4.1. The sensitivity of hydrometeors to the height of freezing level

4.1.1. The development of hydrometeors

When the temperature (at 1000 hPa, $T_{00\text{hPa}}$) changes from 291 K to 279 K, the freezing level is lowered from about 2.85 km to 0.9 km. The vertical cross-sections of mixing ratios of liquid water (summation of cloud water and rain water) and ice water (including ice crystals, snow and graupel) are shown in Figure 3. As airflow approaches the terrain, it is forced to rise. The vertical component of the motion produces an isolated cloud on the windward side of the mountain, and the cloud is evaporated on the lee side. Orographic clouds in these cases are similar in appearance. However, the ice water content increases and becomes more important in the development of orographic cloud with lower freezing levels. When the temperature at 1000 hPa is 291 K ($T_{00\text{hPa}}=291 \text{ K}$), corresponding to the freezing level of 2.85 km ($FL=2.85 \text{ km}$), the ice water content is almost zero and the liquid water content is the largest (Figure 4). The averaged mixing ratios of liquid water in case C_291, case C_288, case C_285, and case C_282 are about 224%, 177%, 112% and 45% larger than that in case C_279 while the averaged mixing ratios of ice water in these cases are about 97%, 78%, 54%, and 23% lower than that in case C_279, or in other words,

liquid water decreases and ice water increases with lowering freezing level. This is consistent with Colle (2004), who reported that rain water decreased while the mixing ratios of snow and graupel increased as the freezing level was lowered from 500 hPa to 1000 hPa. It is also noticed that the mixing ratio of snow is larger than the mixing ratios of ice crystals and graupel in the ice-phase dominated cases (C_282, C_285, and C_288). Case C_291 is composed mainly of liquid water.

4.1.2. The evolution of microphysical processes

Figure 5 shows the time- and grid-integrated mass concentration of water drops contributed by the microphysical processes. The processes of condensation and melting are the main source of the mass of drops while evaporation and riming are the main sink of the mass of drops. When the freezing level is lowered from 2.85 km to 0.9 km ($T_{00\text{hPa}}$ changes from 291 K to 279 K), the growth of drops by condensation becomes less efficient due to completion of available water vapor from ice-phase particles and accretion of drops through various drop-ice interactions (such as riming). Additionally, melting of ice-phase particles into liquid drops gradually becomes effective. For example, the loss of drops by riming process (collection by ice-phase particles) is increased with lowering freezing level. The riming process is the main sink of liquid drops while melting is the main source for the mass of drops (Fig. 5d, e). But the growth of the mass concentration of drops from melting becomes less efficient when the freezing level is lower than 1.9 km or the surface temperature closes to zero. In addition, the processes of riming and melting are dominant in different size ranges of drops. Melting is mainly dominant in the drops large than 100 μm in diameter, especially the drops larger than 300 μm , making an important contribution to precipitation and collection process. However, riming process is mainly dominant in the drops with a diameter less than 100 μm , especially the droplets smaller than 50 μm . When the freezing level is lowered from 2.85 km to 1.9 km, the growth of drops larger than 300 μm by melting process is enhanced, although the loss of smaller droplets by riming process also becomes more significant as the freezing level is lowered.

The development of the ice-phase particles in the cloud (ice crystals, snow, and graupel) is shown in Fig. 6. The growth of ice-phase particles by vapor deposition, including ice depositional growth with expense of liquid droplets (the so-called Wegener–Bergeron–Findeisen process, WBF), becomes more efficient with lowering freezing level. However, due to the growth of mass concentration of ice particles by the WBF process is much less efficient than the growth by riming, the WBF process will not be distinguished from deposition and evaporation in the following discussions. In addition, the formation of ice crystals by the Hallet-Mossop process is much less significant as compared to other primary ice-forming microphysical processes (riming and deposition), the Hallet-Mossop process will not be shown as a stand-alone process. When the freezing level is lowered, the growth of ice water by riming becomes more efficient, especially the growth of snow particles. However, the growth through riming is slowed as the freezing level is lowered from 1.4 km to 0.9km ($T_{00hPa}=282$ K to 279 K), since the available liquid water becomes smaller. The loss of ice water by melting process increases with freezing level lowering from 2.85 km to 1.4 km, and decreases when the freezing level is further lowered. These facts reflect the balance of the effects of temperature and water vapor. On one hand, when the freezing level lows, ice-phase microphysical processes become more important for the formation of precipitation, and more precipitating ice particles are formed through riming in the cloud, leading to more melting. On the other hand, lowering the freezing level also means to a reduction of the available water inside the cloud, so to the ice water content. In the cases simulated here, the height of the freezing level at 1.4 km is a watershed for the importance of riming and melting to raindrops.

4.1.3. Surface precipitation

Figure 7 shows the accumulated precipitation amount and the differences in surface precipitation distribution between the polluted and clean cases. As the freezing level is lowered, the total precipitation amount increases and the percentage of solid precipitation (ice crystals, snow and graupel) along mountain slope also increases. The surface precipitation amount in case C_279 ($FL=0.9$ km) is about 137% more

than that in case C_291 ($FL=2.85$ km). In case C_279, solid precipitation approximately accounts for 85 percent of the total precipitation amount. When the freezing level is elevated to 1.4 km (case C_282), it is approximately 55 percent of the total amount. Moreover, the direct contribution of solid precipitation to surface precipitation amount becomes less important as the freezing level is higher than 1.4 km. As shown in Fig. 4, the predominant precipitation particles in cases C_279 and C_282 ($FL=0.9$ km and 1.4 km) are snow particles while in the other cases, precipitation appears in the forms of rain. This shows again the importance of riming and melting to the surface precipitation.

The results that the sensitivities of hydrometeors and microphysical processes to the height of freezing level, are consistent with Colle and Zeng (2004) and Miglietta and Rotunno (2006), although different thermodynamic profiles are used in previous studies (Miglietta and Rotunno, 2006). The ice-phase microphysical processes, such as riming and deposition, become more efficient in this study as the freezing level is lowered. Although the model and initial setting used in Colle and Zeng (2004) are different from this study, both studies indicate that the accretion of cloud droplets by snow and deposition growth of snow particles became more efficient with lowering the freezing level from 500 hPa to 1000 hPa.

4.2 The sensitivity to increases in aerosol particle concentration

4.2.1. The development of hydrometeors

In order to investigate the effect of aerosols on development of cloud particles and precipitation, the number concentration and size distribution of aerosol particles are changed from a clean continental condition to a polluted urban condition. As aerosol particle number concentration is higher, the liquid water content (summation of cloud water and rain water) under the polluted condition becomes higher than that under the clean condition. Additionally, the ice water content (including ice crystals, snow and graupel) is also significantly higher in the polluted case as compared with the clean case (Fig. 3). However, the responses of cloud water (liquid droplets less than 100 μm in diameter) and rain water to changes in aerosol are different (Fig. 4c, d). Compared

to the clean case, the mixing ratio of cloud water is higher in the polluted case while the mixing ratio of rain water is lower. As we have known, higher aerosol concentration can lead to an increase in number concentration of cloud droplets and a decrease in average droplet size, resulting in more cloud water and less rain (Lynn et al., 2007, Xiao et al., 2014). Hence, it is understandable that the cloud water in the polluted cases is higher than that in the clean cases, and the rain water is lower in the polluted cases. Overall, although the mass of rain water is higher than that of cloud water, the changes in rain water in response to increase in aerosol particle concentration are much less pronounced than that of cloud water. The increase in aerosol particle concentration can delay precipitation by suppressing the collision-coalescence process in warm-phase microphysical processes (Pruppacher and Klett, 1997; Yin et al., 2000) or by enhancing the ice-phase microphysical processes (Lynn et al., 2007; Xiao et al., 2015), leading to a shift in precipitation distribution at the surface. In addition, as the freezing level is higher than 0.9 km, snow is the predominant ice-phase particles and the mixing ratio of snow in the polluted case is higher than that in the clean case.

4.2.2. The evolution of microphysical processes

As the aerosol number concentration is increased, more droplets are activated, resulting in more droplets with a diameter less than 30 μm in the polluted cases, especially the droplets of 10-30 μm in diameter (Figure 8). Hence, the growth of drops of 10-30 μm in diameter by condensation process is enhanced in the polluted cases. The larger amount of droplets with diameters of 10-30 μm promote the growth of drops by collision and coalescence, leading to more mass of drops larger than 300 μm in diameter (Fig. 5c). In addition, the small droplets are easier to be transported to higher levels in the polluted case, leading to higher cloud water content along the vertical direction (Figure 9). Xiao et al. (2015) showed that the small droplets transported to the higher levels could enhance the deposition growth of ice crystals. Pitter and Pruppacher (1974) suggested that the efficiency of riming process increases significantly as the diameter of drops is larger than 10 μm . The larger concentration of

droplets in the size range of 10-30 μm in diameter in the polluted case enhance the growth of solid particles by riming process, especially the growth of snow and graupel particles, leading to more ice water and then more melted water (Fig. 6c, d).

4.2.3. Formation and distribution of precipitation

As the aerosol number concentration is increased, the total precipitation amount in the polluted cases is larger than that in the clean cases (Fig. 7a). Under the conditions with temperature of 279 K and 282 K at 1000 hPa ($FL=0.9$ km and 1.4 km, respectively), the differences in surface precipitation amount between the clean cases and the polluted cases are mainly caused by the changes in solid precipitation, and in other cases with higher freezing levels, changes in liquid precipitation dominate. It is also clear that varying the concentration of aerosol particles leads to a significant change to the distribution of precipitation at the surface (Fig. 7b). Compared to the clean cases, the polluted cases produce less precipitation within the distance of 180 km and more precipitation from 190 km to 210 km along the horizontal axis. Under the polluted condition, more droplets are activated, suppressing the collision-coalescence processes and the formation of large drops, and leading to a shift of drop size distribution to smaller size range and a delay of precipitation at the surface. In addition, as pointed out above, in the polluted cases, more cloud liquid water content mainly lies in the diameter of 10-30 μm which promotes the growth of drops larger than 300 μm by collision and coalescence processes. Hence, there is less precipitation at the early stage but more precipitation along the mountain slope. Because the predominant precipitation contributed by snow in the cases with $T_{00\text{hPa}}$ of 279 K and 282 K, the differences in precipitation distribution induced by increasing aerosol particles are different from other cases with higher freezing levels. However, it is still discernable that the distribution of precipitation at the surface is changed due to increased aerosol particles (Fig. 7b). The larger concentration of small droplets in the polluted cases also promote the growth of ice-phase particles by riming process, especially the growth of snow, leading to more ice water and surface precipitation.

4.3 The sensitivity of aerosol effects to the height of freezing level

4.3.1 The variations of hydrometeors and microphysical processes

When the freezing level is lowered from 2.85 km to 1.9 km ($T_{00hPa} = 291$ K to 285 K), more liquid water is consumed by the growth of ice-phase particles, resulting in less liquid water and less significant change of liquid water to increase in aerosols (Fig. 4). However, the changes in ice water to increasing aerosol concentration become more significant. A more detailed analysis shows that, as the freezing level is lowered, more ice crystals are produced through various ice nucleation mechanisms including condensation-freezing and deposition (not shown). In addition, more small droplets in the polluted cases benefit the growth of ice crystal through the WBF process (Xiao et al., 2015), leading to higher mass of ice-phase particles. In addition, the growth of ice-phase particles by riming is also enhanced and the rate of riming growth caused by increase in aerosols is also promoted with lowering freezing level from 2.85 km to 1.9 km. As a result, the production of rain drops by melting is increased with increase in aerosol particles.

When the freezing level lowers from 1.4 km to 0.9 km, liquid water becomes even smaller, and the change in liquid water with increase in aerosol particle concentration also becomes less significant. However, the riming process plays a dominating role in all these more polluted cases. Compared to the clean cases, the domain-integrated mass concentration of ice-phase particles produced by riming process is about 508 g kg⁻¹, 1454 g kg⁻¹, 1479g kg⁻¹, 1987 g kg⁻¹, and 994 g kg⁻¹ larger in the polluted cases P_291, P_288, P_285, P_282, and P_279, respectively, than the corresponding clean cases. Although the growth of snow particles by riming becomes more efficient with lowering freezing level under polluted condition, there is a reduction in the changes of riming growth to increase in aerosols as the freezing level is lowering from 1.4 km to 0.9 km because the liquid water is reduced as indicated at the end of Section 4.1.2.

4.3.2. The changes of surface precipitation

With increased number concentration of aerosol particles or CCN, the total

precipitation amounts in the polluted cases increase by 5-14% with the freezing level being lowered from 2.85 km to 0.9 km. As the freezing level is lowered from 2.85 km to 1.9 km (the predominant precipitation particles are liquid drops), the changes in precipitation in response to increase in aerosol number concentration is significant and then becomes less significant with further lowering freezing level. The reason is, when the freezing level is lowered from 2.85 km to 1.9 km, more liquid water is consumed for the growth of ice-phase particles. As mentioned above, because less liquid water is produced with lowering freezing level, the aerosol effects on surface precipitation by warm-phase processes (delaying the formation of large drop) become weaker. However, the ice water content increases and the ice-phase processes also become efficient with lowering freezing level. The riming process is more effective in the polluted case, providing large drops by melting and enhancing precipitation. Hence, the surface precipitation is increased with increasing aerosol number concentration. In addition, the mass of drops with diameter larger than 300 μm is also increased by collecting small drops. Although the liquid water produced by melting is increased with lowering freezing level and increase in aerosol particle concentration, the collection of small drops is suppressed due to the shorter distance between the melting level and the ground. Hence, the increase of precipitation caused by increase in aerosol increases firstly and then decreases with further lowering freezing level.

When the freezing level is lowered from 1.4 km to 0.9 km, the ice water content becomes even larger. The major precipitation particle is changed into solid particles and snow particle is the main precipitation particle. When the freezing level is lowered, the percentage of surface solid rainfall to total rainfall is increased, but aerosol particles can still affect the solid particles by influencing liquid drops as discussed above. Riming process is the main source of ice water and the growth of ice water by riming is increased by increase in aerosols, leading to a maximum precipitation enhancement caused by aerosols in the cases of $FL=1.4$ km. In addition, because of the limited liquid water in the cases of $FL=0.9$ km, the change of total precipitation amount in response to increase in aerosols with $FL=0.9$ km is less significant than that with $FL=1.4$ km. Higher aerosol number concentration

suppresses the formation of large drops, and then resulting in delayed precipitation. However, because solid particles become the main precipitation source, the suppression of precipitation at the foot of mountain, which is mainly related to warm-phase process, become less important with lowering freezing level.

5. Summary and conclusions

The responses of mixed-phase orographic cloud and precipitation to changes in environmental concentration and size distribution of aerosol particles under different initial thermodynamic profiles have been analyzed using the Weather Research Forecast (WRF) coupled with a detailed bin microphysics scheme. A mixed-phase orographic cloud developed along an idealized bell-shaped mountain (the peak is at 1.5 km) has been conducted, with a special effort being paid to the sensitivity of aerosol's impact on cloud and precipitation to the altitude of freezing level. The main results of this study are summarized below.

When the freezing level is lowered from about 2.85 km to about 0.9 km ($T_{00hPa}=291$ K to 279 K), the percentage of solid precipitation (ice-phase particles) to total precipitation is increased from zero to about 85% and the mixing ratio of ice water is also increased. When number concentration of aerosols serving as CCN is increased, surface precipitation is increased, but the change of surface precipitation to increase in aerosols is not linear as the freezing level is lowered. The maximum increase of precipitation caused by aerosols is about 14% when the freezing level is at 1.4 km ($T_{00hPa}=282$ K). From the microphysics point of view, this is because more droplets with diameters of 10-30 μm are produced under polluted conditions which promote the growth of ice water by riming process, especially the growth of snow particles, leading to more ice water and benefiting precipitation. Moreover, the change of riming growth to increase in aerosols also becomes more efficient with lowering freezing level, except for the cases with $FL=0.9$ km.

When the freezing level is lowered from 2.85 km to 1.9 km (predominant precipitation particle is liquid drops), the ice water is increased and the change of ice water to increase in aerosol also becomes more significant. The growth of ice particles

by riming is enhanced and the riming growth caused by increase in aerosols is also promoted with lowering freezing level, leading to more ice water for melting. But the distance between the melting level and the ground is decreased with lowering freezing level, suppressing the growth of large drops by collision and coalescence processes. Hence, the change of precipitation to increase in aerosols becomes significant firstly and then less significant with lowering freezing level.

When the freezing level is further lowered ($FL=1.4$ km and 0.9 km), the ice-phase particles play a more important role and the melting process becomes weaker, leading to more ice-phase precipitation. The growth of ice-phase particles by riming also increases with lowering freezing level, but because of the decrease in liquid water, the growth of ice water by riming is suppressed when the freezing level is lowered to 0.9 km ($T_{00hPa}=279$ K), indicating that 1.4 km may be the optimum height of freezing level for the maximum effects of aerosol on precipitation.

Compared to our previous studies (Xiao et al., 2014; 2015), this study provided more cases to investigate the responses of microphysical processes and surface precipitation to the changes in aerosol number concentration under thermodynamic conditions. This study also indicates that the ice-phase particles and their related microphysical processes are important to investigate the effect of aerosol on orographic cloud and that an increase in aerosol particle concentration or CCN leads to a shift in orographic precipitation. This is consistent with previous studies (e.g., Lynn et al., 2007; Saleeby et al., 2013). However, inconsistency exists between the current study and Saleeby et al. (2013) on the effect of aerosols on precipitation. Saleeby et al. showed that increasing CCN concentration reduced the averaged droplet size to less than $10\text{ }\mu\text{m}$ in diameter, and significantly suppressed the efficiency of riming. However, in this study, more droplets with diameters of $10\text{--}30\text{ }\mu\text{m}$ are produced in the polluted case, which actually promote the formation of large drops through collision-coalescence and precipitation particles through riming. Apart from the differences in model details between these two studies, such as the difference between spectral bin microphysics schemes and bulk parameterization schemes (Khain, 2015), the differences in the thermodynamic profiles used in the two studies

may be the main reason for the different results. It seems that the cloud simulated by Saleeby et al. (2013) in a less moist (relative humidity less than 80% before cloud formation) environment and without continuously available vapor for cloud formation. Both the work by Fan et al. (2014) and the present work are based on spectral bin microphysics scheme and the results are consistent with each other. This study provides more insight into the evolution of particle spectra and the changes of aerosol effect under different freezing level.

In addition, the change of precipitation measured from rain gauge which is fixed on one place over the mountain slope, may be the impact of the shift of orographic precipitation or the effect of ice water under different freezing level, leading to an uncertainty of aerosol effect lying at the statistical analyses of rain gauge data (Givati and Rosenfeld, 2004; Rosenfeld and Givati, 2006; Alpert et al., 2008; Halfon et al., 2009). Precipitation may also be influenced by salt, dust, and black carbon in real environment. The rain gauge data may contain other cloud types, such as convective cloud, and may be influenced by complicated weather situations, making the aerosol effects even more complicated (Rosenfeld, 2000; Rosenfeld et al., 2008; Khain, 2009; Tao et al., 2012; Carrió and Cotton, 2014). Hence, further studies are needed to investigate aerosol effects including other factors (such as radiation processes, terrain-blocked effect, aerosol chemical component, etc.) with a detail microphysics scheme.

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666 Table 1. Initial conditions of idealized simulation.

Experiment	Aerosol spectrum	Temperature at 1000 hPa	Approximate freezing level (Above 1000 hPa)
C_279	Clean background	$T_{00\text{hPa}}=279\text{ K}$	$FL=0.9\text{ km}$
P_279	Polluted background		
C_282	Clean background	$T_{00\text{hPa}}=282\text{ K}$	$FL=1.4\text{ km}$
P_282	Polluted background		
C_285	Clean background	$T_{00\text{hPa}}=285\text{ K}$	$FL=1.9\text{ km}$
P_285	Polluted background		
C_288	Clean background	$T_{00\text{hPa}}=288\text{ K}$	$FL=2.35\text{ km}$
P_288	Polluted background		
C_291	Clean background	$T_{00\text{hPa}}=291\text{ K}$	$FL=2.85\text{ km}$
P_291	Polluted background		

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