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Examination of the impacts of ice nuclei aerosol particles on microphysics, precipitation and electrification in a 1.5D aerosol-cloud bin model

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ABSTRACT

To study the impact of dust aerosol particles through their role as ice nuclei (IN) on the development of cloud microphysics and electrification, as well as on their contribution to the precipitation formation, we used a 1.5D detailed microphysics model to conduct sensitivity analysis of the Cooperative Convective Precipitation Experiment (CCOPE) case on 19 July 1981 in Miles City, Montana, USA. The simulated cloud microphysical properties demonstrate that the increase in IN concentrations enhances the number of ice particles produced by heterogeneous nucleation. As a result of increased ice particle formation, the enhanced ice growth via deposition in the Wegener-Bergeron-Findeisen mechanism, more extensive riming and thus an enhancement of ice aggregation, is primarily responsible for the increased numbers of large ice particles. The increased IN concentrations could result in earlier (~6.5 minutes) and stronger (by a factor of 60) precipitation and greater raindrop mass due to enhanced ice phase process. The changes in microphysical processes resulting from increased IN concentrations lead to more large ice particles, which is primarily responsible for the enhanced charge separation process. Additionally, the charge density rises with the increased large ice particle concentrations and both of them reach their maxima at the same height.

1. Introduction

As an important feature in convective clouds, the total lightning activity, including the propagation directions, discharge types and strike locations, is closely related to the charge structures of thunderstorms (e.g., Coleman, Stolzenburg, Marshall, & Stanley, 2008; Guo et al., 2017; MacGorman et al., 2011, 2008; Mansell et al., 2013, 2010; Tan, Tao, Liang, & Zhu, 2014). The thunderstorm charge structure is not only the product of electrification processes, but also is closely associated with the development of microphysical processes in the cloud, especially the ice phase process (e.g., Mitzeva, Saunders, & Tsenova, 2005, 2006, 2009; Mansell, Ziegler, & Bruning, 2010, 2013). Over the past decade, cloud-aerosol interactions have been widely recognized as one of the key

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factors influencing cloud properties. A number of numerical and observational studies have been dedicated to quantifying the impacts of aerosol particles on the cloud microphysics and precipitation (e.g., Rosenfeld et al., 2008, 2012; Saleeby, Cotton, Lowenthal, Borys, & Wetzel, 2009, 2013a, 2013b; Stevens & Feingold, 2009; Khain, Rosenfeld, Pokrovsky, Blahak, & Ryzhkov, 2011, 2015, 2016; Wang et al., 2011; Tao, Chen, Li, Wang, & Zhang, 2012; Xiao, Yin, Jin, Chen, & Chen, 2015; Hazra, Chaudhari, Ranalkar, & Chen, 2017). It has been suggested that aerosol particles, acting as cloud condensation nuclei (CCN), affect the electrification process through modification of cloud microphysics (e.g., Mansell & Ziegler, 2013; Shi et al., 2015; Zhao, Yin, & Xiao, 2015; Tan et al., 2017; Thornton, Virts, Holzworth, & Mitchell, 2017). Recently, growing evidence suggested that aerosol particles acting as ice nuclei (IN), can participate in the formation of ice crystals and cause the enormous alternations of cloud parameters and convective intensity, which may significantly modify the electrification process (e.g., Ault et al., 2011; Gonçalves et al., 2012; Tao et al., 2012; Creamean et al., 2013; Fan et al., 2014, 2017; Hiron & Flossmann, 2015; Boose et al., 2016; Diehl & Grützun, 2018).

Some studies revealed that greater concentration aerosol particles lead to more production of small cloud droplets, the change of precipitation through varying collision efficiencies among hydrometeor particles and increases in large ice phase hydrometeor particle concentrations (e.g., Teller & Levin, 2006; van den Heever, Carrio, Cotton, DeMott, & Prenni, 2006; Wang et al., 2011; Tao et al., 2012; Hazra et al., 2017). As the mechanisms of thunderstorm electrification are intrinsically linked to microphysics, aerosol particles may play a significant role in the thunderstorm electrification. Mansell and Ziegler (2013) and Tan et al. (2017) examined aerosol particle effects by testing a wide range of CCN concentrations in a two-moment bulk microphysics model, and they found that increased aerosol particles lead to earlier and stronger electrification, but further increases would weaken the charge separation. It appears that the electrification in thunderstorms is sensitive to CCN concentrations. However, to the best of our knowledge, no attempt has been made to investigate IN effects on the electrification process in cloud models. Various laboratory results supported that collisions between ice crystals and graupel particles in the presence of supercooled water droplets primarily account for the electrification (e.g., Brooks, Saunders, Mitzeva, & Peck, 1997; Jayaratne, Saunders, & Hallett, 1983; Pereyra, Avila, Castellano, & Saunders, 2000; Saunders et al., 1998, 1991; Takahashi, 1978, 1984; Ziegler, MacGorman, Dye, & Ray, 1991). The storm electrification is therefore highly sensitive to ice particles and the IN effects have been suspected to play a crucial role in the electrification process.

The magnitude and location of charge separated during collisions between ice phase particles are associated with the characteristics of ice crystals, which serve as the predominant charged particles in the noninductive charge separation mechanism. Therefore, the aerosol particles, acting as efficient IN, participating in the formation of ice crystals, may have important impacts on the thunderstorm electrification. Although many studies have investigated the effects of aerosol particles acting as CCN on the electrification, the effects of IN have not been analyzed in cloud models in detail because of the complex interactions between particles and clouds and the lack in understanding the mechanism of IN effects on the nucleation.

Additionally, the impact of IN aerosol particles on the precipitation via modifying cloud microphysics and dynamics has been gradually one of the most concerned questions in the cloud physics. Modeling studies on the effects of IN concentrations on the precipitation were presented by investigators (e.g., Teller & Levin, 2006; Muhlbauer & Lohmann, 2009; Fan et al., 2014, 2017; Diehl & Grützun, 2018). Some studies carried out by Teller and Levin (2006) and Diehl and Grützun (2018) showed that increased concentrations of IN can suppress the precipitation. Another supported by Muhlbauer and Lohmann (2009) and Fan et al. (2014, 2017) suggested that increased IN concentrations have an enhancement effect on the precipitation. It was found from the above studies that the impact of IN on precipitation still remains unclear. Therefore, the challenge further urges us to develop insights into the ice nucleation mechanism that links IN with the ice crystal formation.

To examine the effects of IN aerosol particles, ice nucleation parameterizations have to be linked with aerosol particle properties such as number concentrations, surface area. Most ice nucleation parameterizations used in cloud and climate models are formulated as empirical functions of only temperature and/or supersaturation (Cotton, Tripoli, Rauber, & Mulvihill, 1986; Fletcher, 1962, p. 386; Meyers, DeMott, & Cotton, 1992; Young, 1974), which cannot be used to examine the variability of IN concentrations in the atmosphere and they are usually used outside the measured temperature or supersaturation range. For example, Fletcher parameterization (Fletcher, 1962, p. 386) showed that the relationship in the ice crystal concentrations and temperature is usually exponential, which results in the overestimated ice crystal concentrations at lower temperatures but underestimated ice crystal concentrations at warm temperatures. The parameterization of Meyers et al. (1992), which was applied in many cloud models and derived from measurements between $-7\,^{\circ}\mathrm{C}$ and $-20\,^{\circ}\mathrm{C}$, overestimates the ice crystal concentrations when the effective IN concentrations is low, as shown by Prenni et al. (2007) in the Arctic conditions. In recent years, considerable progress has been made in understanding the impact of IN aerosol particles on the nucleation due to the more comprehensive recognition for the ice nucleation. The IN impacts on clouds are mainly via modifying the heterogeneous nucleation, which has been hypothesized to occur in four different modes: condensation, deposition, contact, and immersion (Pruppacher & Klett, 1997; Vali, 1985). New parameterizations or modifications of existing parameterizations have been made to link aerosol particles with the ice nucleation and in these parameterizations the ice nucleation is dependent on many factors such as temperature, supersaturation and aerosol particle properties (e.g., Barahona & Nenes, 2009; DeMott et al., 2010; Muhlbauer & Lohmann, 2009; Niemand et al., 2012; Phillips et al., 2013, 2008). This recent progress allows us to investigate the IN effects on the precipitation and electrification.

In this study, we numerically investigated how IN (by ice nucleating) impacts the electrification process microphysically and dynamically. The bulk models are too simple to fully detailed describe the particle spectrum evolution and the interaction between aerosol particles and clouds since bulk microphysics cannot microphysically describe the evolution of the IN-containing cloud droplet spectra properly (e.g. Fan, Wang, Rosenfeld, & Liu, 2016; Khain et al., 2015). Additionally, the bulk models usually utilize average fall velocities over the particle spectrum distribution for sedimentation process that does not account for smaller particles which fall slower and larger particles which fall faster, as reviewed by Khain et al. (2015). To solve these problems, we used a two-cylinder time-dependent cloud and aerosol particles interaction model which incorporated the explicit noninductive electrification scheme and spectral bin microphysical scheme. Not only were the number and mass concentrations of aerosol particles, including IN and CCN, explicitly followed in all microphysical processes, but also the charge amounts for a pair of particles in their collisions were determined.

2. The model introduction

2.1. Dynamical framework

To study the microphysical processes of IN, not only is the high resolution of the particle size bins necessary to adequately describe the distributions of aerosol particles (CCN and IN), water droplets and ice particles, but also the mass of CCN or IN in hydrometeors should be followed. We had to choose a one-and-half-dimensional Eulerian model due to current limited computational resources. The first 1.5D cylindrical model was presented by Asai and Kasahara (1967) to investigate the influence of the compensating downward motions on formation and evolution of cumulus cloud. The model geometry is axisymmetric and composed of an outer and an inner cylinder. In such a model, two circular concentric air columns describe the updraft/cloud region (inside column) and the compensating downward motion region (outside column). We further assumed that horizontally, the vertical wind velocities were distributed sinusoidally in order to allow for a vortex-like circulation at the cloud top (Zhao & Austin, 2005). The models of the similar type were used by many researchers to study the effects of aerosol particles on cloud properties (Hiron & Flossmann, 2015; Leroy, Monier, Wobrock, & Flossmann, 2006; Simmel, Bühl, Ansmann, & Tegen, 2015; Sun et al. 2012a,b). For example, Leroy et al., 2006 and Sun et al. 2012a,b studied the effects of aerosol particles as CCN on ice and warm clouds, respectively. In other models of the similar type, the radii ratio of the two cylinders was set to be 10 (Hiron & Flossmann, 2015; Leroy et al., 2006; Simmel et al., 2015). We performed a number of sensitivity tests and compared model results with observations of the Cooperative Convective Precipitation Experiment (CCOPE; Dye et al., 1986). It was found that the radii ratio of 4 (r_a = 3.2 km, r_b = 12.8 km, respectively) leads to a good agreement between the simulated the cloud top rise rate and the maximum cloud top height and the observations (Sun et al. 2012a,b). Fig. 1 shows the dynamics of the model. The lateral exchanges between the cylinders are allowed via the horizontal velocity $\tilde{u}$. The vertical wind velocities are $\bar{w}_a$ and $\bar{w}_b$. The vertical resolution was chosen to be 100 m. The time step of 2s was adopted in the dynamics which was the same as in the microphysics.

2.2. Microphysics

The microphysical processes especially focus on the interactions between clouds and aerosol particles in the model. In the atmosphere, Aerosol particle properties are much more complicated than we describe in cloud models. In order to simplify the situation, only one kind of IN and CCN is considered in our study. Aerosol particles, water droplets and ice particles are described by two three-dimensional number density distribution functions $f_{wat}(m, m_{AP}, x_{sulfate/IN})$ and $f_{ice}(m, m_{AP}, x_{sulfate/IN})$, where m is the hydrometeor mass, m_{AP} is the mass of the aerosol particles that serve as CCN and IN, and x represents two species of aerosol particles. One of these is CCN, while the other is IN, such as bacteria, pollen, or dust. We assume that CCN and IN are composed of ammonium sulfates and dust, respectively. In this model, the sizes of growing cloud droplets and raindrops are determined by condensation,

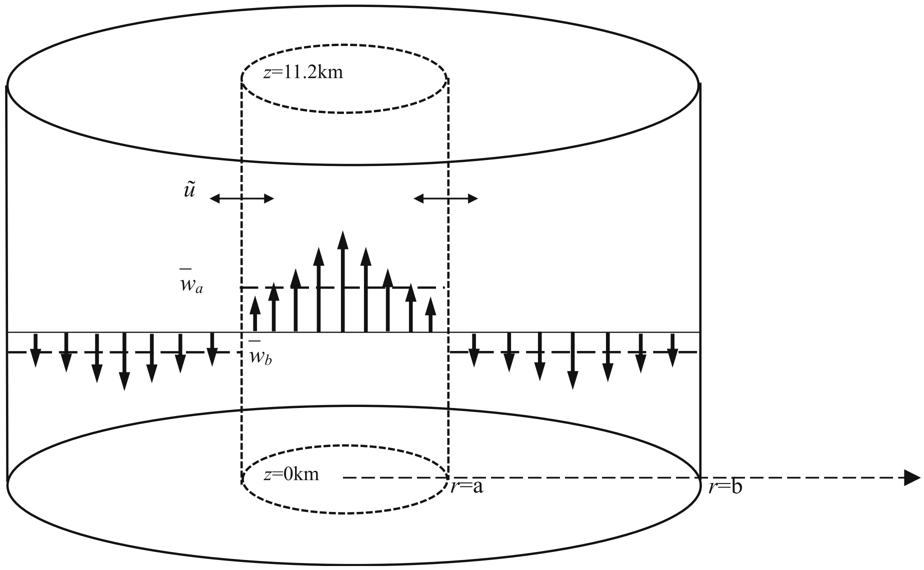

Fig. 1. Schematic view of the 1.5D model with updraft region (inside column, r_a = 3.2 km) and subsidence region (outside-inside column, r_b-r_a; r_b = 12.8 km).

evaporation and stochastic collision and coalescence (e.g., Leroy et al., 2006; Sun, 2008; 2012a, 2012b). The ice phase microphysics includes nucleation, multiplication, deposition, aggregation, riming, and melting. Many studies (e.g., Leroy et al., 2006; Sun, 2008; 2012a,b; Flossmann and Wobrock, 2010; Hiron & Flossmann, 2015) described the liquid phase and ice phase processes in detail and give the detailed formula of the parameterized schemes. Therefore, and most of them will not be repeated here. It should be noted that similar to the melting process in the 1.5D framework by Hiron and Flossmann (2015), it is assumed that the ice particles can melt completely and instantaneously when they reach the 0 °C level in the model. In the numerical calculations, the logarithmic mass grid is equally spaced with 90 bins for the mass of aerosol particles in radius from 8.0×10^{-3} μm to 2.36×10^{2} μm and 130 bins for the mass of hydrometeors in radius from 8.0×10^{-3} μm to 2.4×10^{4} μm in the mixing phase cloud study. In this approach, the categories with the smallest size are interpreted as ice crystals (radius $<$ 200 μm), while larger ice particles are considered as graupel (radius $\geq$ 200 μm). Droplets are divided into cloud droplets (radius $<$ 50 μm) and raindrops (radius $\geq$ 50 μm) (Sun, 2008; 2012a,b).

In order to study the possible effects of IN aerosol particles on the thunderstorm microphysics, precipitation, and electrification process, an empirical parameterization of heterogeneous ice nucleation proposed by Phillips, DeMott, and Andronache (2008) has been implemented in the 1.5D cloud model. The parameterization is based on deposition, condensation, and immersion freezing modes. It represents the known and empirically quantified modes of heterogeneous ice nucleation. The number of active IN is determined by supersaturation with respect to ice, air temperature, nucleation threshold temperature, proportion of this kind of IN in the total IN concentrations, and size and concentration distributions of IN aerosol particles. IN species are grouped by the parameterization into three basic types: dust particles, soot/black carbon, and bioaerosols represented by bacteria. Thus, the proportion of active dust used in 1.5D model needs to be determined when implementing the parameterization. It has been assumed that the proportional contribution of this IN type is a constant value that is independent of humidity and temperature (see Phillips et al. (2008) and Sun et al. (2012a,b)). The relative contribution of the dust spectrum used in the study was set at 2/3 (Rogers & DeMott, 2001a; 2001b). Details of the parameterization and how to implement it were noted by Phillips et al. (2008).

2.3. Electrification

Laboratory studies have suggested that rebounding collisions between ice crystals and graupel particles can result in an appreciable separation charge in the presence of supercooled water droplets. This process is referred to as the noninductive charge separation. It is generally considered as the main charge separation mechanism in the thunderstorm electrification (e.g. Mansell et al., 2013, 2010; Shi et al., 2015; Tan et al., 2017; Zhao et al., 2015). Similar to Mansell et al. (2010, 2013) and Shi et al. (2015), only noninductive charge separation was adopted in the study to simulate the storm electrification. Over the past several decades, it has been widely recognized that the significant charge is transferred when ice crystals rebound from graupel particles (e.g., Brooks et al., 1997; Jayaratne et al., 1983; Mansell & Ziegler, 2013; Saunders et al., 1998, 1991; Shi et al., 2015; Takahashi, 1978; Ziegler et al., 1991). In our study, this collision occurs between small and large ice particles. The model uses one size distribution function to describe hydrometeors. The ice particles have 130 bins with the radius from 8.0×10^{-3} to 2.4×10^{4} μm. In this approach, the small ice particles are ice crystals with the radius of less than 200 μm and large ice particles are graupel particles with the radius of greater than 200 μm. The general formulation for the noninductive graupel charging separation rate (Q_t) takes the form:

$$Q_t = K_{gi} \cdot n_g \cdot n_i \cdot \delta q \tag{1}$$

$$K_{gi} = \frac{\pi}{4}(D_i + D_g)^2 \Delta V_{gi} \varepsilon_{gi} \tag{2}$$

where, K_{gi} is the collision separation kernel, n_i and n_g are the number concentrations of ice crystal and graupel, and δq is the charge separated per rebounding collision between graupel and ice crystal. The particles that collide with each other obtain an equal amount of opposite charge. D_i and D_g are the diameters of the collision particles (ice crystal and graupel). ΔV_{gi} is the relative velocity of colliding particles and ε_{gi} is the collision separation rate.

The scheme of individual graupel charging separation rate (δq) used was Saunders et al.'s scheme (defined as S98) (Saunders and Peck, 1998) which was based on a number of previous experimental results (Jayaratne et al., 1983; Jayaratne & Saunders, 1985; Keith & Saunders, 1989, 1990; Brooks et al., 1997; Saunders and Peck, 1998). δq is in S98 scheme given by:

$$\delta q = k_q D_i^{m} |V_g - V_i|^{n} q(R) \tag{3}$$

where k_q, m, and n are constants that depend on the ice crystal size as shown in Table 1. D_i is the ice crystal diameter, V_i and V_g is the terminal fall velocity of ice crystal and graupel. $q(R)$ is calculated from Eq. (5) and Eq. (6) below. The polarity of charge gained by

Table 1
Values of Constants k_q, m, and n used in equation (1) (from Brooks et al., (1997)).

Charge sign	Crystal size (μm)	k_q	m	n
+	< 155	4.92×10^{13}	3.76	2.5
+	155–452	4.04×10^{13}	1.9	2.5
+	> 452	52.8	0.44	2.5
–	< 253	5.24×10^{8}	2.54	2.8
–	≥253	24	0.5	2.8

graupel is closely related to the rime accretion rate R ($R = ELWC \times V_g$, where ELWC is the effective liquid water content which is experienced by the riming graupel pellets in the cloud).

Saunders and Peck (1998) defined positive and negative graupel charging zones by constructing a curve of critical R_{crit} as a function of temperature. In dependence of the difference between R and R_{crit} graupel obtains a negative or positive charge at a particular temperature. The critical rime accretion rate R_{crit} is given by:

$$R_{crit} = \begin{cases} s(T) & : T > -23.7\,°\mathrm{C} \\ k(T) & : -23.7 > T > -32.47\,°\mathrm{C} \\ 0 & : T \leq -32.47\,°\mathrm{C} \end{cases} \tag{4}$$

where $s(T)$ and $k(T)$ is the function of temperature.

The positive charge acquired by graupel is determined by ($R > R_{crit}$):

$$q(R) = 6.74(R - R_{crit}) \tag{5}$$

and the negative charge acquired by graupel is determined by ($0.1\mathrm{g{\cdot}m^{-2}{\cdot}s^{-1}}R < R_{crit}$):

$$q(R) = 3.9(R_{crit} - 0.1) \times \left(4 \times \left[\frac{R - \frac{(R_{crit} + 0.1)}{2}}{(R_{crit} - 0.1)}\right]^2 - 1\right) \tag{6}$$

The discharge process refers to the work by Zhao et al., 2015 in the 1.5D model. It is a simple overall discharge parameterization scheme and the simplified process is quite different from the real discharge process (Details were noted by Zhao et al., 2015). Additionally, the focus of this study is the overall response of electrification process to the IN concentrations, and the magnitude of charge in simulated thunderclouds is one of the most important points to be investigated. It should be point that the transferred charge only obtained from the electrification is unrelated to the lightning discharge. Therefore, the discharge process is only for the purpose of decline of charge and it is not discussed in detail.

2.4. Background field description

The initial conditions for the temperature and humidity fields used the profiles measured during the Cooperative Convective Precipitation Experiment (CCOPE) on 19 July 1981 in Miles City, Montana, USA (Dye et al., 1986). One of the aims of the CCOPE was to study the relationship between the development of microphysical and dynamic processes and the electrical evolution in cumulus clouds. Under moderately unstable stratification and relatively weak wind shear, an isolated convective cloud developed. In this study we assumed that IN was composed of dust and CCN was composed of ammonium sulfate. This aerosol spectrum was fitted by superimposing three log-normal distribution functions and the aerosol distribution may be approximated by (Yin, Levin, Reisin, & Tzivion, 2000):

$$\frac{dN}{dlnr_n} = \sum_{i=1}^{3} \frac{n_i}{(2\pi)^{1/2} log\sigma_i ln10} \exp\left(-\frac{[log(r_n/(R_i)]^2}{2(log\sigma_i)^2}\right) \tag{7}$$

where n_i is the particle number concentration in mode i, R_i the geometric mean radius in mode i, σ_i the geometric standard deviation, and r_n the radius of aerosol particles. These distribution parameters were similar to those used by O'Dowd et al. (1997) and the initial aerosol spectrum parameters as described in Eq. (7) is given in Tables 2 and 3. N is the total particle number. The total sulfate aerosol particles in all cases were characterized by parameters $N = 1236\,\mathrm{cm}^{-3}$ (Fig. 2). Because slight changes in IN concentrations can significantly affect cloud properties, the total numbers of dust IN were set $N = 0.3\,\mathrm{cm}^{-3}$ (case 1), $0.8\,\mathrm{cm}^{-3}$ (case 2), and $1.3\,\mathrm{cm}^{-3}$ (case 3), respectively, based on the measurements by Rogers and DeMott (2001a; 2001b) and the work by Sun (2008). The IN spectra were uni-modal (Tong & Lightart, 2000) as shown in Fig. 3, with parameters listed in Table 3.

The background aerosol particle spectrum was assumed to decrease with height. Therefore, the concentration can be expressed as Yin et al. (2000):

$$N(z) = N(z = 0) * \exp\left(-\frac{z}{z_s}\right) \tag{8}$$

where z_s (the scale height) was set to 2 km in this study. $N(z)$ is the aerosol particle concentration at the ground.

Table 2
Parameters for the CCN distribution: n_i (cm^{-3}) is the particle number concentration in mode i, R_i (μm) is the geometric mean radius in mode i, σ_i is the geometric standard deviation.

Mode i	n_i	R_i	σ_i
1	798	0.03	1.9
2	420	0.1	1.9
3	18	1.0	2.0

Table 3
Parameters for the IN distribution: n (cm^{-3}) is the particle number concentration, R (μm) is the geometric mean radius, σ is the geometric standard deviation.

Case	n	R	σ
1	0.3	0.4	2.0
2	0.8	0.4	2.0
3	1.3	0.4	2.0

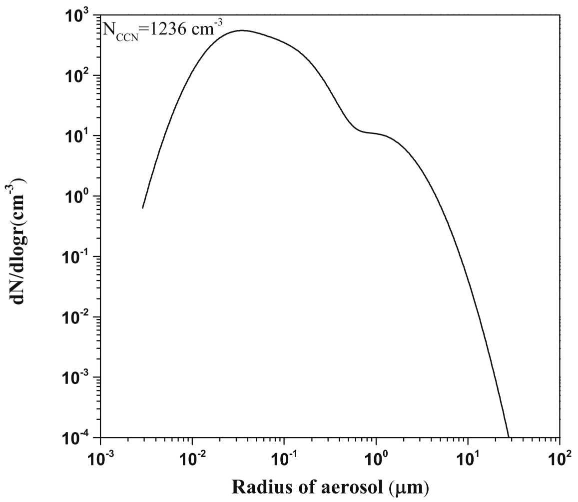

Fig. 2. CCN distribution in three cases.

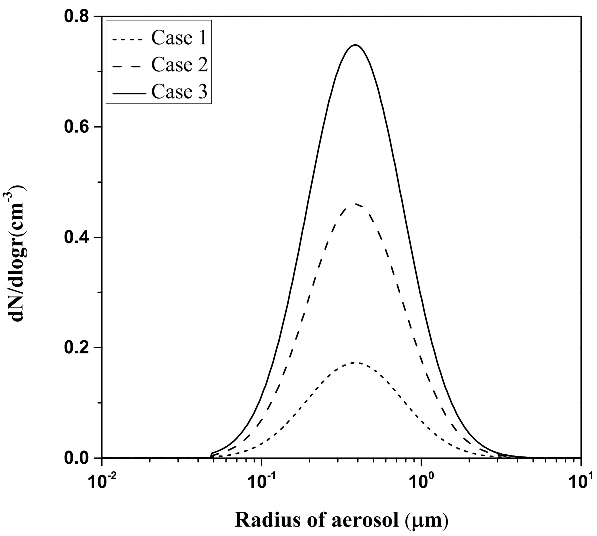

Fig. 3. IN distribution in three cases.

3. Results

3.1. IN impact on the microphysics and dynamics

The simulation results of the dynamics and microphysical process have already been compared to the CCOPE measurements in the 1.5D framework by Leroy et al. (2006) and based on the framework of their study, Hiron and Flossmann (2015) used the observation of CCOPE to simulate the ice initiation in the thunderstorm. Therefore, most of them not be repeated here. In the framework, only the variation of the results because of changes in the IN concentrations shall be studied.

According to earlier studies, the changes of IN concentrations not only affect the development of microphysical processes, but also

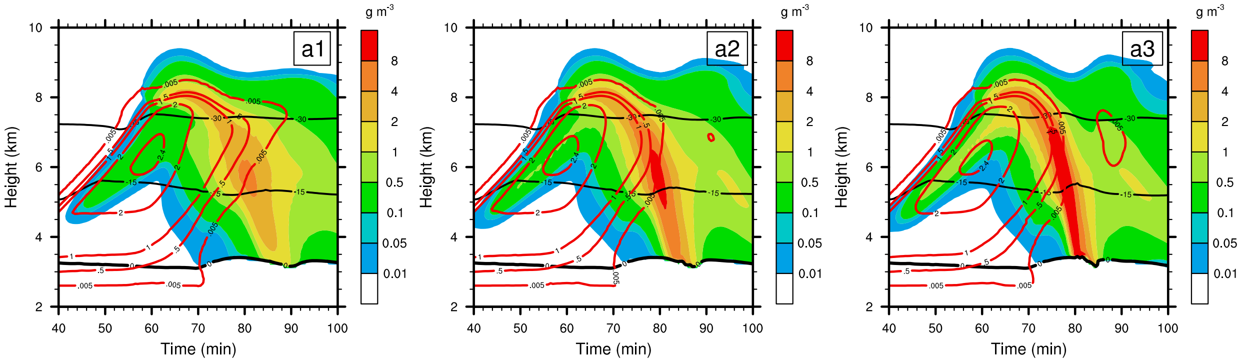

Fig. 4. Evolution of the cloud droplet mass (red lines) and ice mass (color areas) for three cases. The black lines mark the temperature levels of −30, −15, and 0 °C. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

the dynamical structure in clouds (e.g., Teller & Levin, 2006; van den Heever et al., 2006; Ekman, Engstrom, & Wang, 2007; Tao et al., 2012; Fan et al., 2014, 2016, 2017). In order to confirm these findings and, thus, to prove the performance of our model, three cases with different IN concentrations as described in Section 2.4 were simulated. Fig. 4 displays the time and height evolution of cloud droplet mass and ice mass (including ice crystals and graupel particles) for each case. Cloud droplets are mainly distributed at height of 1.5–8.5 km and ice starts to form on above 4.2 km altitude at about 40 min. The comparison between the drop and ice mass indicates that the changes in the IN concentrations not only impact mass distribution of the cloud droplets especially in the mature stage of ice phase (60–90 min) but also have a greater impact on the whole ice phase process. The consumption of cloud droplets in case 3 is more intense than those in case 1 and 2. The ice phase process exhibits the expected sensitivity to the increased IN concentrations, yielding greater ice mass as concluded by Fan et al. (2014, 2017). A detailed comparison of ice crystals and graupel mass shows both ice crystal and graupel mass are enhanced with increased IN (Fig. 5). It is obvious from Table 4 that the maximum graupel mass is 4.4 g m^{-3}, 11.5 g m^{-3} and 23.6 g m^{-3} and the maximum ice crystals mass is 2.1 g m^{-3}, 2.5 g m^{-3} and 4.0 g m^{-3} in case 1, 2 and 3, respectively.

Higher IN concentrations could directly enhance the heterogeneous nucleation, leading to producing ice crystals more efficiently. Once formed, the ice crystals continue to grow by vapor deposition, riming and aggregation to form large ice particles. Fig. 6 shows the number size distribution of ice particles larger than 115 μm in radius. From Fig. 6, it can be seen that an increase in IN concentrations contributes to enhance concentrations of the large ice particles. The increase in IN concentrations directly enhanced the amount of ice produced by heterogeneous nucleation, resulting in enhanced ice growth via deposition in the Wegener-Bergeron-Findeisen mechanism (i.e., ice particles grow at the expense of liquid particles due to the lower saturation vapor pressure over ice compared with that over liquid at a given temperature). Besides, by riming, ice particles can also be converted into large ice particles, especially graupel particles. Although enhanced ice formation reduces liquid particles available for riming, riming can occur more extensively in a liquid rich condition due to increased ice particle concentrations. The increase in the number of large ice particles via deposition and riming as a result of increased IN concentrations, then facilitates aggregation process. As expected, these enhanced ice processes cause a faster conversion of liquid phase to ice phase in the cloud. Consequently, as a result of the increased IN concentrations, these processes strengthen the consumption of supercooled water which is the main reason for decrease in the distribution of cloud droplets in the mature stage of ice phase (60–90 min) (Fig. 4).

In addition to affecting microphysical properties, variations in IN concentrations also have the potential to change the dynamics of the storm system according to earlier studies (e.g., Teller & Levin, 2006; van den Heever et al., 2006; Ekman et al., 2007; Tao et al., 2012; Fan et al., 2014, 2017). This is illustrated in Fig. 7. In particular, the updrafts reach higher altitudes in case 3, compared with

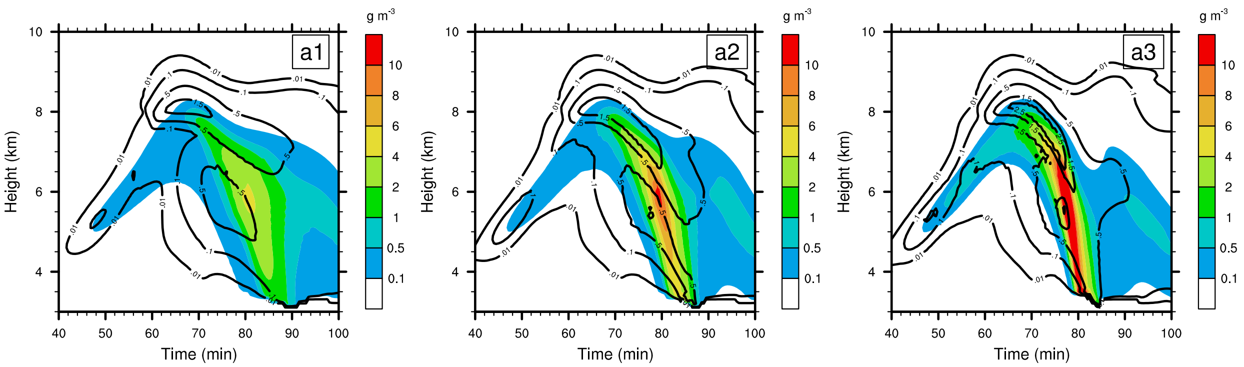

Fig. 5. Evolution of the ice crystals mass (solid line) and graupel mass (color area) for three case. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

Table 4
The simulated maximum values and corresponding time and height from three cases.

Feature	Case 1	Case 2	Case 3
Max. LWC (g m^{-3})/time and height (min, km)	2.6/57,6.4	7.1/84,3.2	19.1/81,3.1
Max. cloud droplet mass (g m^{-3})/time and height (min, km)	2.5/59,6.3	2.5/59,6.4	2.4/59,6.3
Max. raindrop mass (g m^{-3})/time and height (min, km)	1.9/87,3.0	7.1/84,3.2	19.1/81,3.1
Max. ice crystal number (cm^{-3})/time and height (min, km)	66.2/58,8.2	73.8/58,8.2	74.7/58,8.2
Max. graupel number (L^{-1})/time and height (min, km)	43.4/80,6.1	85.4/79,6.2	150.8/77,6.2
Max. ice crystal mass (g m^{-3})/time and height (min, km)	2.1/66,8.1	2.5/66,8.1	4.0/75,7.1
Max. graupel mass (g m^{-3})/time and height (min, km)	4.4/80,6.1	11.5/78,6.0	23.6/77,5.8
Max. precipitation rate (mm h^{-1})/time (min)	3.1/92	38.1/89	182.5/85
Max. ice crystal charge density (nc m^{-3})/time and height (min, km) (absolute value)	1.3/71,7.8	1.5/70,7.8	1.9/69,7.8
Max. graupel charge density (nc m^{-3})/time and height (min, km) (absolute value)	1.1/71,7.6	1.25/69,7.7	1.5/68,7.7
Max. total charge density (nc m^{-3})/time and height (min, km) (absolute value)	0.8/72,7.8	1.0/71,7.9	1.2/70,7.9

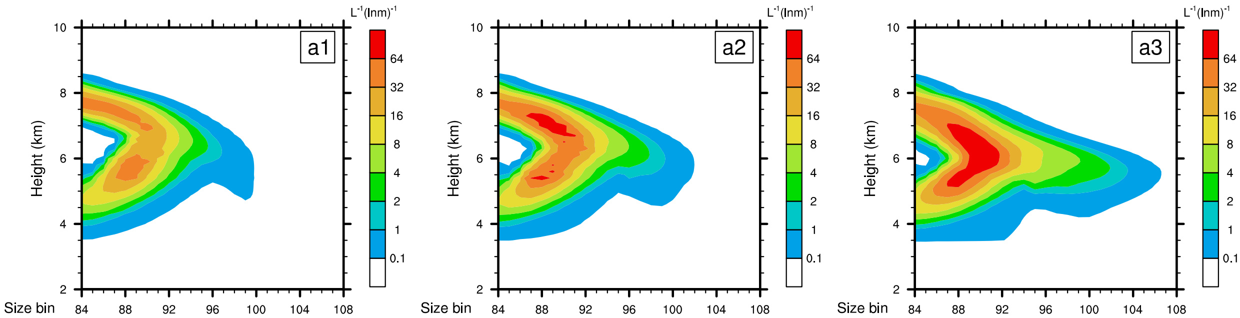

Fig. 6. Number concentrations of large ice particles (r > 115 μm) as function of particle radius at 75 min for three cases.

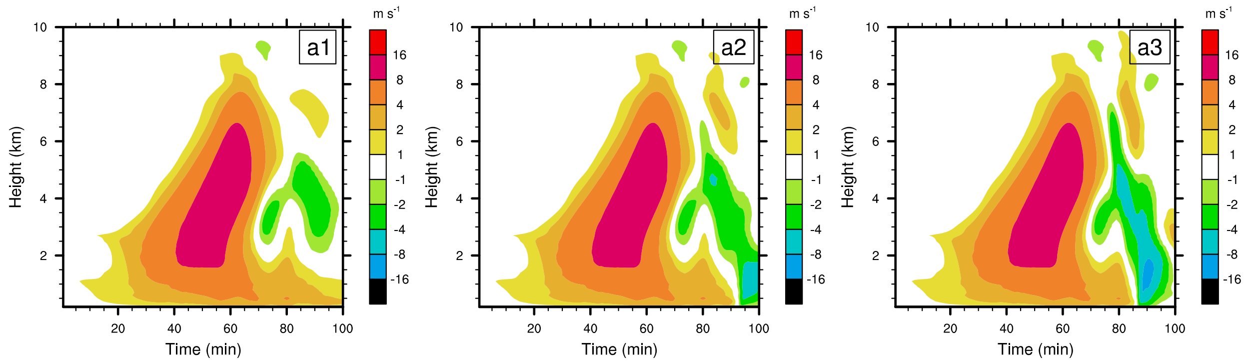

Fig. 7. Evolution of the vertical velocity (m s^{-1}) for three cases.

the lower concentration cases (case 2 and 3) during the dissipating storm stages. The cause of the changes in dynamical structure is that increased aerosol particle concentrations are conducive to the formation of greater ice mass, leading to greater associated release of latent heat upon interacting with other hydrometeors. This finding is consistent with studies of van den Heever et al. (2006) and Ekman et al. (2007). Stronger updrafts can enhance the upward transport of supercooled water which further enhances the formation and growth of ice particles at the cloud top during the dissipating storm stages (compare Fig. 4-a1 and a2; Fig. 4-a1 and a3). Obviously, these enhanced ice phase processes in turn lead to the release of greater amounts of latent heat which subsequently increase the updraft strength. This results finally in an enhancing feedback loop.

3.2. IN impact on the precipitation

To study the impact of aerosol particles as CCN on the precipitation formation, Leroy et al. (2006) conducted a number of sensitivity studies of the 19th July 1981 CCOPE case in the 1.5D framework and main microphysical characteristic was the formation of precipitation mainly via the ice phase. In their study ice particles formed by heterogeneous nucleation was described by an empirical function of Meyers et al. (1992) and the nucleation parameterization used by Leroy et al. (2006) was not linked with properties of IN aerosol particles. Fig. 4 illustrates that the increased IN concentrations cause a faster conversion of liquid to ice in the

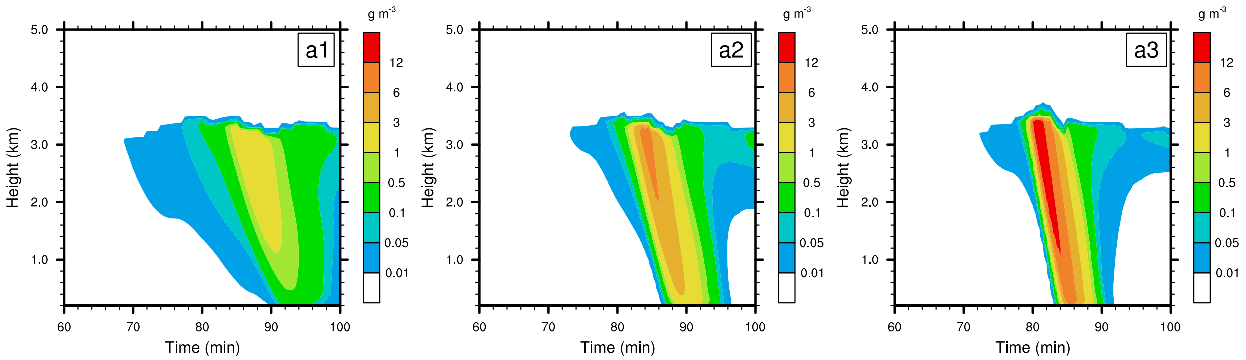

Fig. 8. Evolution of raindrop mass for three cases.

mixed-phase cloud. Therefore, although the heterogeneous nucleation parameterizations used in these two models are different, the precipitation in our study is also formed mainly via the ice phase. The rain initiation is defined as the time at which the maximum precipitation rate at the surface begins to exceed 0.1 mm h^{-1}. Fig. 8 shows that raindrops start to form after about 70 min and their masses are enhanced as IN concentrations increase (maximum mass is 1.9 g m^{-3}, 7.1 g m^{-3} and 19.1 g m^{-3} in case 1, 2 and 3, respectively). Simultaneously, Fig. 9 shows that increased IN concentrations result in higher precipitation rates and earlier onset of precipitation. The increase of the IN concentration from 0.3 to 0.8 cm-3 (cases 1 and 2) enhances the maximum precipitation rate from 3.1 to 38.1 mm h^{-1} while with the highest IN concentration (case 3) the maximum precipitation rate reaches 182.1 mm h^{-1} (Table 4 and Fig. 9). Therefore, the precipitation rate has increased by about 60 times and the precipitation has been advanced by 5 minutes. As shown in Section 3.1, higher IN concentrations increase the ice mass. The formed ice particles grow once they collide with drops and form large ice particles which fall and melt on the way to the ground. It is assumed that the ice particles can melt completely and instantaneously when they reach the 0 °C level in the model. Consequently, more precipitation can result from the melting of large ice particles. The conclusion can be found in the work by Sun (2008).

3.3. IN impact on the electrification process

The noninductive charging mechanism is extremely sensitive to the ice phase process (e.g., Mitzeva, Latham, & Petrova, 2006, 2009, 2005; Mansell et al., 2013, 2010; Shi et al., 2015; Tan et al., 2017; Thornton et al., 2017; Zhao et al., 2015). Considering that the alternations of ice phase processes with varying IN concentrations will significantly affect the electrification since the non-inductive charge separation depends on ice particle collisions (Saunders and Peck, 1998).

During the entire development in the thunderstorm the maximum total charge density is 0.8 nC m^{-3} in case 1, 1.0 nC m^{-3} in case 2, and 1.2 nC m^{-3} in case 3 (Table 4). Additionally, as a result of increased IN concentrations, the total charge density acquired by ice crystals and graupel particles is significantly enhanced (Table 4). According to S98 scheme, it is an expected result that the electrification process increases as more ice crystals and graupel particles are generated.

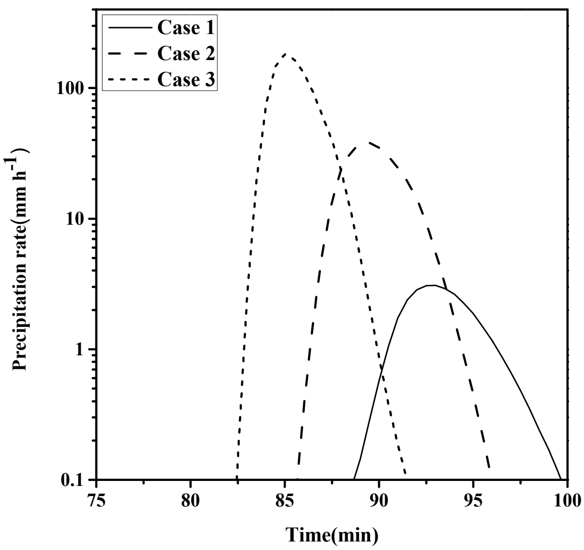

Fig. 9. Evolution of Precipitation rate for three cases.

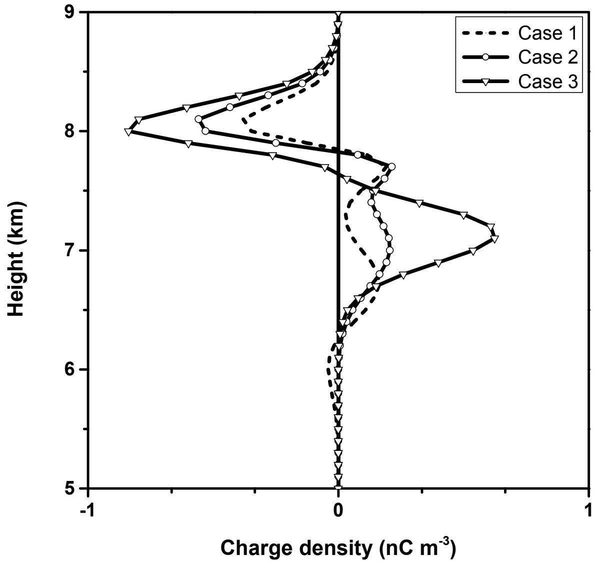

Fig. 10. Total charge density at t = 67 min for three cases.

Fig. 10 shows that during the early (pre-lightning) stage of thunderstorm electrification (at t = 67 min) the three cases have a similar charge structure, developing an inverted dipole charge structure consisting of a net positive charge region at 6.5–7.7 km altitude, a net negative charge above 7.7–8.7 km. The peak value of charge density at t = 67 min in case 3 is about 0.8 nC m^{-3}, stronger than those in case 1, 0.6 nC m^{-3} and case 2, 0.4 nC m^{-3}. Consequently, increased IN concentrations can enhance the total charge density. Fig. 11 displays the evolution of large ice particle concentrations (r > 140 μm) as a function of height at t = 67 min. Comparing Fig. 10 with Fig. 11, it is evident that at height of 7.7–8.0 km, the charge density rises with the increased large ice particle concentrations and both of them reach their maxima at ~8.0 km. It can be found that the dramatic increase of large ice particles generated in microphysical processes is primarily responsible for the enhanced charge separation process in the corresponding region.

In order to explore the electrification of ice particles more clearly, Fig. 12 illustrates the impact of IN concentrations on the net charge density dominated by each bin charged particles in the early stages of electrification, when the discharge did not occur. We can see that there is a dramatic increase of the net charge density acquired by different mass-binned ice crystal and graupel with the increased IN concentrations. According to S98 scheme, the increase of charged particle concentrations, especially the large ice particles, can significantly enhance the charge separation process. Therefore, it is not difficult to conclude that increased IN concentrations can enhance the charge density acquired by different mass-binned ice particles due to increased large mass-binned ice

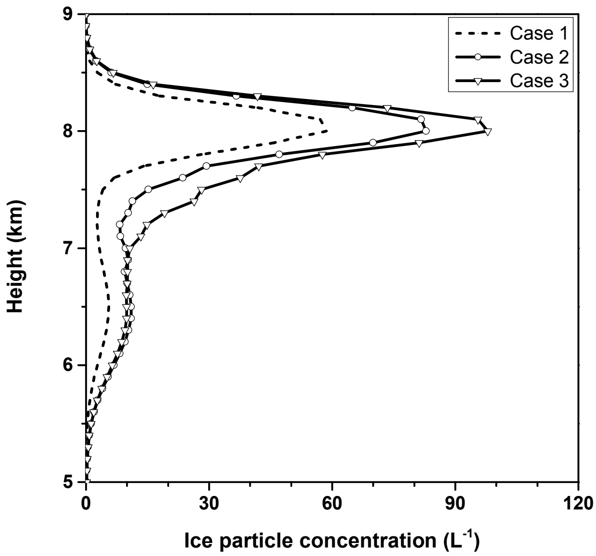

Fig. 11. The large ice particle concentration (r > 140 μm) at t = 67 min for three cases.

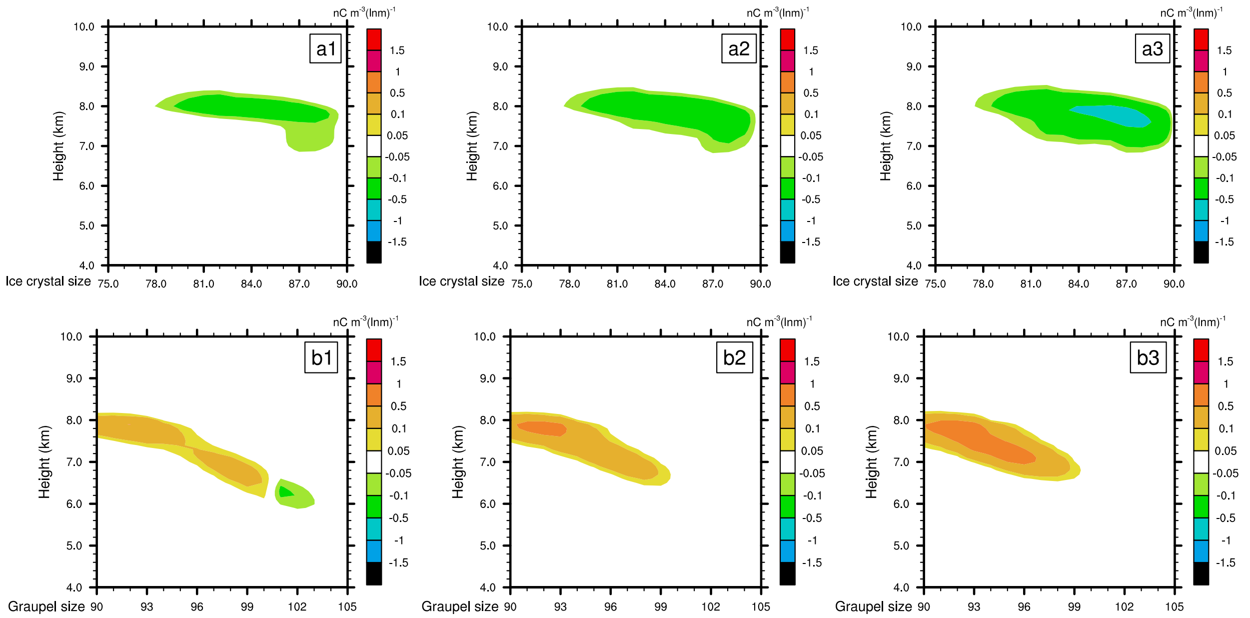

Fig. 12. The charge density varies at t = 67 min for three cases, upper panel: ice crystals, lower panel: graupel.

particles.

4. Conclusion

A 1.5D cylindrical model has been employed to investigate the development of a thunderstorm. A convective cloud event occurring on 19 July 1981 in Miles City, Montana, USA, was simulated and the model results were used to explore the impact of dust through their role as IN on thunderstorm properties by changing the initial dust concentrations. The analysis of the results demonstrated a strong dependency of the development of cloud on the initial IN concentrations. The most important findings are the following:

Under the same meteorological conditions, variations in aerosol particle concentrations not only affect microphysical properties, but also have the potential to change the dynamics of the storm system. The increase in IN concentrations enhances the amount of ice particles produced by heterogeneous nucleation, thus promoting the ice growth via deposition in the Wegener-Bergeron-Findeisen mechanism, more extensive riming and thus an enhancement of ice aggregation, which is primarily responsible for the increased numbers of large ice particles. Due to the enhancement of the ice phase process, the increased latent heat release associated strengthens updrafts during the dissipating stages. The addition of updrafts, in turn, increases the ice mass at the cloud top.

The increase of IN concentrations from 0.3 to 1.3 cm^{-3} results in earlier (~6.5 minutes) and stronger (by a factor of 60) precipitation due to the formation of more precipitation-sized ice particles, considering the fact that the greatest contribution to precipitation formation from ice phase process. It is found that the charge density acquired by different mass-binned ice particles and total charge density acquired by all ice particles are enhanced as large ice particles increase with elevated IN concentrations. Additionally, the charge density rises with the increased large ice particle concentrations and both of them reach their maxima at the same height. Therefore, in the atmosphere, IN concentrations play a primary role in modifying cloud properties and electrification process. The IN effects should be considered in cloud models.

The 1.5D model includes a detailed description of sophisticated ice process. The distributions of aerosol particles and hydrometeors are described by high resolution of the particle spectrums. The model includes impacts of IN aerosol particles on the warm rain and ice formation. Therefore, this study not only provides a better understanding of IN effects on mixed-phase cloud properties and precipitation, but may also be used to better understand the ice phase processes that rely heavily on the drop size spectrum that can lead to extreme lightning events.

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