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Aerosol effects on the development of deep convective clouds using a detailed CRM

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1 Introduction

Changes in ambient concentrations of cloud condensation nuclei (CCN) and ice nuclei (IN) potentially alter cloud properties that may ultimately lead to modifications in cloud radiative forcing and/or precipitation. Considerable attention has been given to the effects of aerosol particles on cloud properties for warm stratiform clouds (e.g., Ackerman et al., 2004; Lu and Seinfeld, 2006; Sandu et al., 2008; Hill et al., 2008, 2009; Wang and Feingold, 2009a,b; Wang et al., 2010). The extent to which these processes hold in mixed-phase and/or cold clouds is not well established. The ice phase presents significant complexities not present in warm clouds (i.e., riming, aggregation, accretion, heterogeneous and homogeneous freezing, melting, etc.), and the cold-rain process is the predominant mechanism by which rain forms (not collision-coalescence of liquid drops). Recently, the potential effects of polluted environments on the formation and development of deep convective clouds has received attention via both modeling studies (e.g., Koren et al., 2005; Van den Hoever et al., 2006; Van den Hoever and Cotton, 2007; Khain et al., 2008; Rosenfeld et al., 2008; Stevens and Feingold, 2009; Khain and Lynn, 2009; Fan et al., 2009) and, less commonly, observational analyses (e.g., Koren et al., 2010).

Conceptual hypotheses have been put forth by Rosenfeld et al. (2008) and Stevens and Feingold (2009) for the invigoration of deep convective clouds by increased aerosol loading. Via different reasoning, both works conclude that an increase in aerosol number concentration should act to increase surface precipitation. Although this makes sense conceptually, modeling studies are still not in agreement as to the sign of the effect on precipitation owing to increased pollutants. For example, Van den Hoever et al. (2006) showed that adding aerosol particles in the form of CCN, giant CCN (GCCN), and/or IN caused a decrease in domain-average cumulative precipitation in reference to a clean environment observed during the Cirrus Regional Study of Tropical Anvils and Cirrus Layer-Florida Area Cirrus Experiment (CRYSTAL-FACE). On the other hand, Khain and Lynn (2009) demonstrated an increase in precipitation with an increase in CCN concentration using a spectral bin microphysics model but with low spatial resolution and minimal simulation time. In the same study, a decrease in precipitation with an increase in CCN number concentration was shown using a simple two-moment bulk microphysics scheme.

Measurements of IN number concentration were performed during CRYSTAL-FACE within a period of enhanced dust particle concentration (DeMott et al., 2003; Sassen et al., 2003). DeMott et al. (2003) reported that during CRYSTAL-FACE, IN number concentrations were observed to be as high as 1 cm^{-3} ($10^3\ \ell^{-1}$). Later, Van den Hoever et al. (2006) and Teller and Levin (2006) demonstrated a decrease in precipitation with an increase in IN concentration using 3D and 2D CRMs, respectively. However, these studies do not fully represent the potential effects of IN on deep convective cloud development since the freezing process is parameterized based on the empirical relation of Meyers et al. (1992) in which the IN number concentration is expressed as an exponential function of temperature and/or supersaturation. For low temperatures (i.e., less than about -30°C), the IN

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number concentration becomes erroneously large and will likely significantly impact the model predictions.

Microphysical calculations of deep convective cloud invigoration in response to aerosol changes have been performed in recent years (e.g., Khain et al., 2004; Teller and Levin, 2006; Khain et al., 2008; Khain and Lynn, 2009). Potential shortcomings exist in the method in which the CCN concentration is implemented and the representation of the IN number concentration in many studies employ the empirical Twomey relationship to predict the number of activated aerosol particles as a function of supersaturation (Twomey, 1959). The empirical constants in this relation are specific to individual cloud types, i.e., the coefficients that apply for the convective core may not be adequate for other regions of the deep convective cloud, e.g., detrained stratocumulus. Moreover, some of the previous studies have used two-dimensional models (e.g., Khain et al., 2004; Teller and Levin, 2006; Khain et al., 2008) and others that have simulated all three dimensions (e.g., Khain and Lynn, 2009) have been performed at rather low spatial resolution, i.e., >1 km in the horizontal. It is natural to ask if with limited computational resources, should one simulate deep convective clouds using detailed bin microphysics or instead use a detailed two-moment bulk scheme at much higher spatial resolution? And, if one accounts for the activation of cloud droplets and nucleation of ice particles in a more physically coherent manner, what are the effects of aerosol particles on precipitation in deep convective clouds? These points are addressed in this study.

2 Numerical Methods

2.1 Microphysics Scheme

We employ the two-moment bulk microphysics scheme of Morrison et al. (2005) and Morrison and Pinto (2005), included with the Weather Research and Forecasting (WRF) model. The scheme has a fixed cloud drop number concentration (N_c), and the freezing process is parameterized following Cooper (1986). These processes are modified as follows.

2.1.1 Cloud Droplet Activation

We have implemented a state-of-the-art activation scheme following Nenes and Seinfeld (2003) and Fountoukis and Nenes (2005). The scheme allows for a sectional representation of the aerosol size distribution. However, to reduce the computational burden required to predict the number of activated aerosol particles, we assume a single-mode lognormal size distribution. The maximum supersaturation (s_{max}) in a grid cell is estimated by employing the "population splitting" concept of Nenes and Seinfeld (2003) in which the growth of the subsequently formed cloud droplets is split into two categories: (1) those drops that grow significantly beyond their critical size (D_c) and (2) those drops that experience little growth beyond D_c. The number of activated aerosol particles, or, the number of cloud droplets is then computed by integrating the size distribution to get,

$$N_{act} = \int_0^{s_{max}} n^s\left(s'\right) ds' = \frac{N_a}{2} erfc\left[\frac{2\ln\left(\frac{s_g}{s_{max}}\right)}{3\sqrt{2}\ln\sigma}\right] \tag{1}$$

where, $erfc$ is the error function complement, and s_g is the geometric mean critical supersaturation for the critical supersaturation distribution,

$$s_g = \sqrt{\frac{4f_1^3\rho_l M_s}{27i\rho_s M_w D_g^3}}. \tag{2}$$

Here, ρ_s is the density of the solute, respectively, M_w and M_s are the molecular weights of water and the solute, respectively, i is the van't Hoff dissociation factor, and f_1 is defined by,

$$f_1 = \frac{4\sigma_w M_w}{\rho_w} \tag{3}$$

in which σ_w is the surface tension of water. For details on calculating s_{max} see Nenes and Seinfeld (2003) and/or Fountoukis and Nenes (2005).

2.1.2 Homogeneous and Heterogeneous Freezing of Cloud Droplets

Morrison et al. (2005) and Morrison and Pinto (2005) use the parameterization of Cooper (1986) to predict the number of ice nuclei (IN) at a given temperature. Here, we employ the homogeneous and heterogeneous freezing parameterization for a monodisperse IN population of Barahona and Nenes (2008, 2009) to predict the number concentration of ice crystals (N_i). The physical basis of the parameterization comes from an approximate solution to the system of equations that define a parcel of cloudy air undergoing freezing, i.e., changes in supersaturation with respect to ice, ice water mixing ratio, and the subsequent growth of ice crystals after freezing. For more details on the calculations required to compute N_i, see Barahona and Nenes (2009). Given a predicted value of N_i ($N_i\,(t+\Delta t)$), the actual rate of freezing can be computed as

$$\left.\frac{dN_i}{dt}\right|_{frz} = \frac{N_i(t+\Delta t)-N_i(t)}{dt} \iff N_i(t+\Delta t) > N_i(t) \qquad (4)$$

while the rate of change of the ice water mixing ratio due to freezing is straightforward once $dN_i/dt|_{frz}$ has been computed.

2.2 Experimental Setup

The WRF model, modified as described in Section 2, is initialized with an idealized sounding typical for continental locales conducive to deep convective development (Figs. 1 and 2). Two soundings are used in order to analyze the extent to which an aerosol-induced effect on deep convection is dependent upon the ambient moisture content, i.e., the water vapor mixing ratio (q_v) or relative humidity (RH). The ambient RH is permitted to change with height similar to that of Khain and Lynn (2009), except that in the present study, the RH at the surface is 95% in the moist scenarios and the RH for the drier scenarios is simply 10% less than that of the moist cases (hereinafter these scenarios are referred to as the highRH and lowRH simulations, respectively). Therefore, the RH at the surface is 85% for the lowRH simulations. Recently, Fan et al. (2009) showed that aerosol effects may be negligible on deep convective clouds in high shear environments. As a result, we limit the vertical wind shear by utilizing the standard quarter circle shear wind profile derived from Weisman and Klemp (1982) (Fig. 2). Convection is initiated in the domain with a perturbation (bubble) in the potential temperature field.

We use a three-dimensional (3D) domain. The horizontal domain length is 250 km in both the x- and y-direction while the vertical domain extends from the surface to 20 km. This vertical depth allows us to simulate into the lower stratosphere which is important for properly depicting anvil formation near the tropopause. We employ 250 grid points in the horizontal (grid spacing is 1000 m) and 80 grid points in the vertical (mean grid spacing is 250 m) for simulations shown here. A time step of 6 s is used to ensure numerical stability. The duration of the simulations is 8 hr. This duration allows us to capture the point at which the precipitation rate reaches a maximum and then declines to nearly zero.

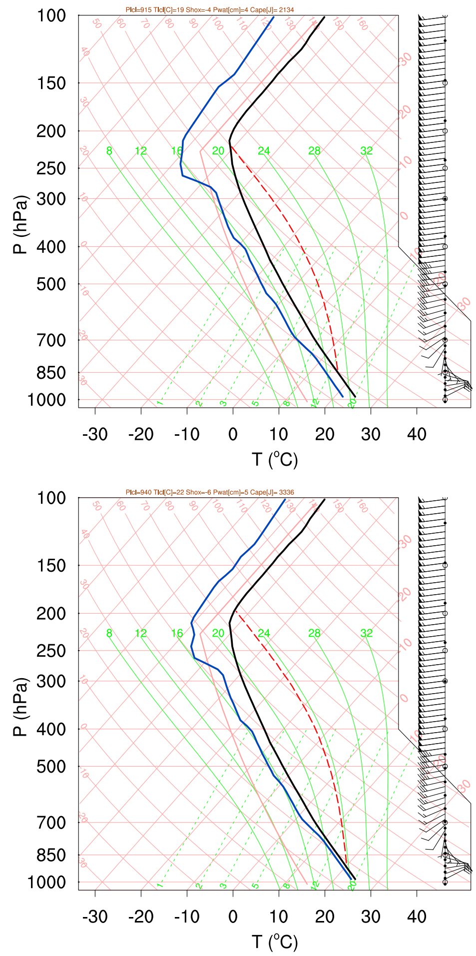

Figure 1: Skew T-Log-P diagrams of the initial temperature and moisture data for the (a) lowRH and (b) highRH simulations. The soundings are adopted from Khain and Lynn (2009) with modifications.

To analyze the potential effects of CCN and IN on deep convective clouds we perform a set of three simulations with varying concentrations of CCN and IN. These simulations are defined as: (1) "Clean" - N_{CCN} = 100 cm^{-3} and N_{IN} = 10 ℓ^{-1}, (2) "Polluted" - N_{CCN} = 1000 cm^{-3} and N_{IN} = 10 ℓ^{-1}, and (3) "IN-Polluted" - N_{CCN} = 1000 cm^{-3} and N_{IN} = 1000 ℓ^{-1}. The "clean" scenario will be used as the base case. The purpose of the "polluted" and "IN-polluted" cases is to show the effect of an increase in aerosol concentration when the particles act only as CCN and when they are CCN and/or IN, respectively.

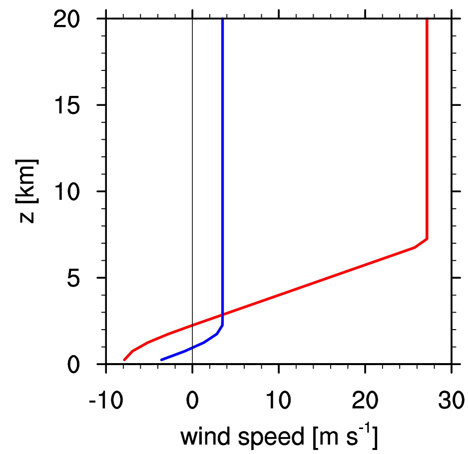

Figure 2: Quarter circle shear wind profile. The zonal wind (u) is in red and and the meridional wind (v) is in blue. The values are derived following Weisman and Klemp (1982) as modified for inclusion in WRF

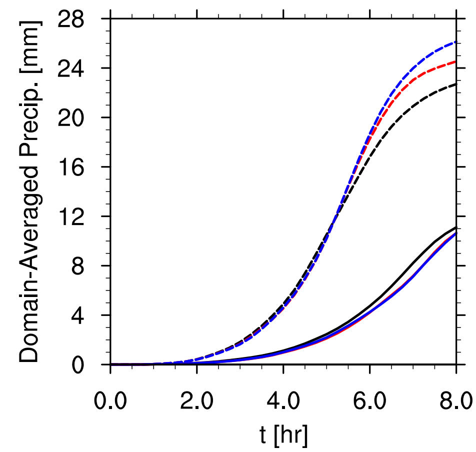

Figure 3: Domain-averaged cumulative precipitation for lowRH (solid) and highRH (dashed) simulations. CCN and IN effects are shown for the "clean" (black), "polluted" (red), and "IN-polluted" (blue) scenarios.

3 Results

3.1 CCN Effect on Precipitation

The overall effect of a perturbation in the CCN number concentration is to modify the precipitation amounting from a deep convective storm cloud. We depict this as the domain-average cumulative surface precipitation in Fig. 3. If the additional aerosol particles act only as CCN, the effect is to increase (decrease) the domain-averaged cumulative precipitation for the highRH (lowRH) simulation by 8.1% (4.2%) as shown in Table 1. Although this RH-dependent aerosol-induced effect has been shown previously using a bin microphysics model (Khain and Lynn, 2009, for example), we demonstrate a qualitatively similar result using a detailed bulk microphysics scheme coupled to a detailed cloud droplet activation scheme (as opposed to assuming a fixed value for N_c or alternatively the Twomey relation for N_c as a function of supersaturation). It is prudent to include a precise and efficient method for calculating the number of nucleated cloud droplets in a bulk microphysics scheme, since autoconversion rate is often proportional to N_c^b, where b is an empirically derived constant. The bulk scheme used here (Morrison et al., 2005; Morrison and Pinto, 2005) makes use of Khairoutdinov and Kogan (2000) in which b is -1.79. Hence, small changes in N_c can have significant effects on the autoconversion process and thus the precipitable water.

We find that in the lowRH simulations, for all levels and all times, there is no apparent tendency for the vertical velocities to be higher in the "polluted" compared to the "clean" case (not shown). On the other hand, the highRH simulations exhibit different signatures, especially during the period in which the domain-averaged precipitation curves diverge (i.e., $5 \leq t < 7$ hr, see Fig. 3). At $t = 5$ hr, there is an increase in the number of grid points with higher vertical velocities when the CCN concentration is elevated, a clear sign of convective invigoration. Then, at $t = 6$ hr, the invigoration of the cloud is concentrated in the upper levels of the cloud, i.e., $z \geq 8.0$ km. This is in agreement with the hypothesis of Stevens and Feingold (2009). The cloud dissipates rapidly from this point on (not shown). The invigoration of the cloud in the upper levels is confirmed by the increased evaporational cooling (shown here as the change in T from the "clean" to "Polluted" and "IN-Polluted" cases in Fig. 4a) and the increase in cloud top height corresponding to an increase in CCN and IN concentrations (Fig. 4b). The end result of such an increase in the strength of the convection is to increase the cumulative precipitation (Fig. 3).

3.2 IN Effect on Precipitation

The effect of an increase in the IN number concentration on deep convective clouds is demonstrated here by use of the homogeneous and heterogeneous freezing parameterization of Barahona and Nenes (2008) and Barahona and Nenes (2009). Fig. 3 shows that when the increased aerosol concentration is assumed to contribute to both an increase in the CCN and

Table 1: Domain-averaged cumulative precipitation at the completion of the simulations performed, i.e., at $t = 8$ hr

RH Profile	"Clean" P	"Polluted" P	"Polluted" ΔP^a	"IN-Polluted" P	"IN-Polluted" ΔP^b
lowRH	11.11 mm	10.64 mm	-4.2%	10.64 mm	-4.2% (0.0%)
highRH	22.70 mm	24.54 mm	8.1%	26.14 mm	15.2% (6.5%)

[a] The relative change in the domain-averaged cumulative precipitation (ΔP) is computed for the "Polluted" case compared with that of the "Clean" case.
[b] ΔP is computed for the "IN-Polluted" case compared with that of the "Clean" case. ΔP between the "IN-Polluted" and "Polluted" cases is given in parentheses.

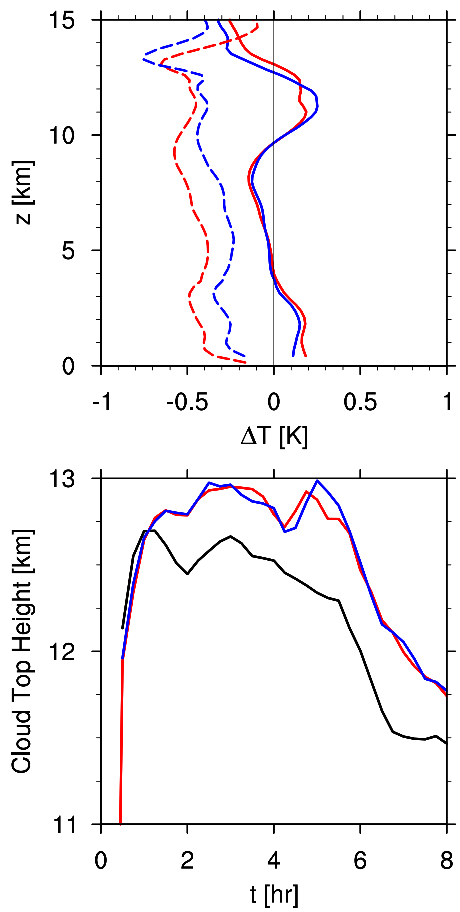

Figure 4: Shown is (a) cloud top height for the "clean" (black), "polluted" (red), and "IN-polluted" (blue) highRH simulations and (b) the change in temperature (ΔT) between simulations. ΔT is depicted for "polluted"-"clean" (red) and "IN-polluted"-"clean" (blue) from the lowRH (solid) and highRH (dashed) scenarios. Cloud top height is defined as the level in which the total condensed water mixing ratio (q_t) is less than 0.01 g kg^{-1}.

the IN number concentrations, the domain-average cumulative precipitation increases by 15.2% ("IN-polluted" case compared to that of the "clean" case, see Table 1). In other words, our simulations show that the effect of an increase in aerosol number concentration is nearly doubled when *both* the CCN *and* IN number concentrations are elevated. From Table 1, the simulations predict an increase in the cumulative precipitation by 6.5% due to an increase in the IN number concentration when compared with the "Polluted" case.

The predicted increase in precipitation due to an increase in IN is based on the fact that an increase in the IN number concentration causes heterogeneous ice nucleation to occur more readily for $T >$ -40°C. As a result, as the condensed hydrometeors are advected upward in an updraft, they are more likely to freeze at warmer temperatures, and thus lower in the atmosphere. These particles are able to grow rapidly since the equilibrium vapor pressure with respect to ice is lower than that of liquid, hence effectively removing water vapor and inhibiting homogeneous freezing (DeMott et al., 1997). As a result, when the IN concentration is relatively low, homogeneous freezing will dominate, while heterogeneous freezing becomes increasingly more important as the IN concentration increases and approaches a homogeneous nucleation limiting concentration (N_{lim}) at which homogeneous freezing is entirely prevented (Barahona and Nenes, 2009). The end result is an increase in N_c and consequently q_c in the mixed-phase region of the cloud. Note that N_c and q_c do not necessarily increase proportionately due to an increase in IN and so one may expect that r_e will change as a result. This increase in N_c and q_c causes the riming process to be enhanced and thus, q_s and q_g increase. The increase in riming owing to an increase in the IN number concentration is most significant before the influx of precipitation (i.e., $t < 5$ hr). This makes sense since time is needed for the particles to fall from the level at which they are formed to reach the freezing level. The enhancement of q_s and q_g then results in an increase in the melting rates below the freezing level and subsequently and increase in q_r and finally the cumulative precipitation.

4 Conclusions

A state-of-the-art aerosol activation parameterization and cloud drop freezing parameterization are implemented into a detailed two-moment mixed-phase bulk microphysics scheme in WRF. Our simulations suggest that an increase in the CCN concentration ("Polluted case) is likely to cause an increase in precipitation by increasing the cloud top height and hence the capacity of the cloud to hold condensed water. We show that an invigoration of the simulated deep convective clouds results from cloud top evaporation/sublimation as opposed to be dominantly controlled by differential latent heating within the cloud itself. Moreover, simulations in which both the IN and CCN concentrations are elevated (IN-polluted) show that the effect of an increase in the aerosol number concentration may be nearly double that of just including CCN effects alone. We relate this further enhancement in precipitation to an invigoration of the riming process that in turn causes an increase in the melting of snow and graupel. The end result is a further increase in the precipitable water and hence more precipitation. These results are dependent upon the ambient moisture content, as it is shown that when the RH decreases by 10%, a decrease in the cumulative precipitation is predicted for an increase in the CCN concentration.

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