Aerosol Impacts on Simulated Supercell Thunderstorms in VORTEX2 and VORTEX-SE

Mingyang Guo
Purdue University

Follow this and additional works at: https://docs.lib.purdue.edu/open_access_theses

Recommended Citation
https://docs.lib.purdue.edu/open_access_theses/1536

This document has been made available through Purdue e-Pubs, a service of the Purdue University Libraries. Please contact epubs@purdue.edu for additional information.
AEROSOL IMPACTS ON SIMULATED SUPERCELL THUNDERSTORMS IN VORTEX2 AND VORTEX-SE

by

Mingyang Guo

A Thesis

Submitted to the Faculty of Purdue University
In Partial Fulfillment of the Requirements for the degree of

Master of Science

Department of Earth, Atmospheric, & Planetary Sciences
West Lafayette, Indiana
August 2018
THE PURDUE UNIVERSITY GRADUATE SCHOOL
STATEMENT OF COMMITTEE APPROVAL

Dr. Robin L. Tanamachi, Chair
   Department of Earth, Atmospheric, & Planetary Sciences
Dr. Daniel T. Dawson, II
   Department of Earth, Atmospheric, & Planetary Sciences
Dr. Michael E. Baldwin
   Department of Earth, Atmospheric, & Planetary Sciences

Approved by:
   Dr. Darryl Granger
   Head of the Graduate Program
In Dedication to Mom and Dad
ACKNOWLEDGMENTS

I would first like to thank my thesis Advisor Dr. Dawson, my committee members Dr. Baldwin and Dr. Tanamachi, and Dr. Mansell from NOAA for help with my research and writing. I would also like to acknowledge my friends who provided help with my coursework and research.

I would also like to thank my boyfriend, Sha Lou, for accompanying me through high and low. He was always there with continuous encouragement when I was confused, distressed and self-doubting.

Finally, I would like to express my great gratitude to my parents, Tong Liu and Tiecheng Guo, for providing me with unfailing love and unconditional support. I would not have achieved this without them.
# TABLE OF CONTENTS

LIST OF TABLES ........................................................................................................................................ VI

LIST OF FIGURES ....................................................................................................................................... VII

ABSTRACT .................................................................................................................................................. X

CHAPTER 1 INTRODUCTION ...................................................................................................................... 1

CHAPTER 2 BACKGROUND .......................................................................................................................... 4

CHAPTER 3 METHODOLOGY ......................................................................................................................... 11

3.1 Model configuration .......................................................................................................................... 11

3.2 Microphysics Scheme ......................................................................................................................... 14

3.3 Budget Analysis .................................................................................................................................. 18

CHAPTER 4 RESULTS ................................................................................................................................ 19

4.1 CCN Impacts on Storm Evolution ...................................................................................................... 19

4.2 CCN Impacts on Hydrometeors ........................................................................................................ 20

4.3 CCN Impacts on Updrafts .................................................................................................................. 31

4.4 CCN Impacts on Precipitation ........................................................................................................... 34

4.5 CCN Impacts on Cold Pools .............................................................................................................. 36

4.6 The 060509 Case .................................................................................................................................. 42

CHAPTER 5 CONCLUSION AND DISCUSSION ......................................................................................... 52

APPENDIX .................................................................................................................................................. 56

REFERENCES .............................................................................................................................................. 58
LIST OF TABLES

Table 1 Convection modes simulated in previous studies.......................................................... 5
Table 2 ARPS options in model simulations................................................................................ 13
Table 3 Sounding parameters........................................................................................................ 14
Table 4 Bulk densities of hydrometeors in the scheme ............................................................... 17
Table 5 Percentage change of cold pool parameters..................................................................... 41
Table 6 Ratio of total graupel sublimation and melting cooling, and rain evaporative cooling... 44
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 1</td>
<td>Skew-T plots and hodographs for (a) 033116, (b) 043016, (c) 060509 and (d) 060909, blue dots mark heights at 1000, 3000, 6000 and 12000 m (if applicable)</td>
<td>12</td>
</tr>
<tr>
<td>Figure 2</td>
<td>Surface radar reflectivity (colored contours) at 60, 90 and 120 min for the 033116 case at CCN100 (a-c) and CCN3000 (d-f)</td>
<td>19</td>
</tr>
<tr>
<td>Figure 3</td>
<td>Same as Figure 2 but for the 043016 case</td>
<td>20</td>
</tr>
<tr>
<td>Figure 4</td>
<td>Same as Figure 2 but for the 060509 case</td>
<td>21</td>
</tr>
<tr>
<td>Figure 5</td>
<td>Same as Figure 2 but for the 060909 case</td>
<td>21</td>
</tr>
<tr>
<td>Figure 6</td>
<td>Surface radar reflectivity (colored contours) at 150 (a-b) and 180 min (c-d) for the 060509 case</td>
<td>22</td>
</tr>
<tr>
<td>Figure 7</td>
<td>Vertical profile of domain-averaged cloud water mass mixing ratio from 60 to 120 min in (a) 033116, (b) 043016, (c) 060509, (d) 060909</td>
<td>23</td>
</tr>
<tr>
<td>Figure 8</td>
<td>Vertical profile of domain-averaged cloud mean mass diameter from 60 to 120 min in (a) 033116, (b) 043016, (c) 060509, (d) 060909</td>
<td>24</td>
</tr>
<tr>
<td>Figure 9</td>
<td>Same as Figure 7 but for rain mass mixing ratio</td>
<td>25</td>
</tr>
<tr>
<td>Figure 10</td>
<td>Same as Figure 8 but for rain mean mass diameter</td>
<td>26</td>
</tr>
<tr>
<td>Figure 11</td>
<td>Same as Figure 7 but for graupel mass mixing ratio</td>
<td>27</td>
</tr>
<tr>
<td>Figure 12</td>
<td>Same as Figure 8 but for graupel mean mass diameter</td>
<td>28</td>
</tr>
<tr>
<td>Figure 13</td>
<td>Same as Figure 7 but for hail mass mixing ratio</td>
<td>29</td>
</tr>
<tr>
<td>Figure 14</td>
<td>Same as Figure 8 but for hail mean mass diameter</td>
<td>29</td>
</tr>
<tr>
<td>Figure 15</td>
<td>Vertical profile of domain-averaged total mass budget in freezing of rain (adgj), graupel accretion of cloud (behk), and graupel accretion of rain (cfil) during 60 to 120 min in 033116 (a-c), 043016 (d-f), 060509 (g-i), and 060909 (j-l)</td>
<td>30</td>
</tr>
<tr>
<td>Figure 16</td>
<td>Mean mass diameter probability density distribution of graupel at 3.5-6 km in (a) 033116, 4-6 km in (b) 043016, 3-6 km in (c) 060509 and 4-8 km in (d) 060909</td>
<td>31</td>
</tr>
<tr>
<td>Figure 17</td>
<td>Same as Figure 16 but for graupel conversion to hail (adgj), hail accretion of cloud (behk), and hail accretion of rain (cfil)</td>
<td>32</td>
</tr>
</tbody>
</table>
Figure 18 Updraft volume of w>10 m s$^{-1}$ (top) and maximum vertical velocity (bottom) over the whole simulation time period in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

Figure 19 Total latent heating released by cloud condensation and freezing over the whole simulation time period in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

Figure 20 Total surface precipitation rate from 20 min to the end of simulation in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

Figure 21 Surface precipitation rate (colored contours) at 60 min at CCN 100 cm$^{-3}$ (a-c) and 3000 cm$^{-3}$ (def), vertical velocity (line contours) at 3 km with levels of 5 m s$^{-1}$ (thin solid) and 15 m s$^{-1}$ (thick solid) and downdraft of -3 m s$^{-1}$ (dashed) in 033116 (ad), 043016 (be), and 060909 (cf).

Figure 22 Same as Figure 21, but at 90 min.

Figure 23 Surface mean mass diameter of rain (ad), graupel (be) and hail (cf) at 60 min at CCN 100 cm$^{-3}$ (a-c) and 3000 cm$^{-3}$ (d-f), vertical velocity (line contours) at 3 km with levels of 5 m s$^{-1}$ (thin solid) and 15 m s$^{-1}$ (thick solid) and downdraft of -3 m s$^{-1}$ (dashed) for 033116.

Figure 24 Same as Figure 23, but at 90 min.

Figure 25 Surface mean mass diameter of (a) rain, (b) graupel and (c) hail at 60 min at CCN 100 cm$^{-3}$ (top) and 3000 cm$^{-3}$ (bottom), vertical velocity (line contours) at 3 km with levels of 5 m s$^{-1}$ (thin solid) and 15 m s$^{-1}$ (thick solid) and downdraft of -3 m s$^{-1}$ (dashed) for 043016.

Figure 26 Same as Figure 25, but at 90 min.

Figure 27 Cold pool area of equivalent potential temperature perturbation ($\theta^e$) < -1 K (top), minimum $\theta^e$ (middle) and average $\theta^e$ in the cold pool (bottom) from 20 min to the end of simulation in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

Figure 28 Surface equivalent potential temperature (colored contours) at 60 min at CCN 100 cm$^{-3}$ and 3000 cm$^{-3}$, and 90 min at CCN 100 cm$^{-3}$ and 3000 cm$^{-3}$, vertical velocity (line contours) at 3 km with levels of 5 m s$^{-1}$ (thin solid) and 15 m s$^{-1}$ (thick solid) and downdraft of -3 m s$^{-1}$ (dashed) in 033116 (a-d), 043016 (e-h), 060509 (i-l), and 060909 (m-p).

Figure 29 Vertical profile of domain averaged rain evaporative cooling from 60 to 120 min in (a) 033116, (b) 043016, (c) 060509, and (d) 060909.

Figure 30 Same as Figure 27 but for graupel sublimation (dashed) and graupel melting (solid).

Figure 31 Vertical profile of domain averaged cloud water and ice mass mixing ratio from 60 to 120 min in (a) 060509 and (b) 060909, and (c) sounding temperature profiles.
Figure 32 As in Figure 1 but for the 060509 case after interpolation onto the 060909 case vertical levels. ................................................................. 45

Figure 33 Surface radar reflectivity (colored contours) at CCN 100 cm$^{-3}$ at 60 min (a-c), 90 min (d-f) and 120 min (g-i) for the 060509rh (adg), 06509uv (beh), and 060509rhuv (cfi). .......... 45

Figure 34 Same as Figure 35 but for CCN3000. ................................................................. 46

Figure 35 Vertical profile of domain averaged mass mixing ratio from 60 to 120 min of (a) cloud water, (b) rain, (c) graupel and (d) hail for 060509uv. ................................................................. 47

Figure 36 Vertical profile of domain averaged evaporation fraction from 60 to 120 min in (a)033116, (b)043016, (c)060509, (d)060909, (e)060509uv, and (f)060509rhuv......................... 48

Figure 37 Updraft volume of $w>10$ m s$^{-1}$ (top) and maximum vertical velocity (bottom) over the whole simulation time period for 060509uv................................................................. 49

Figure 38 Total surface precipitation rate from 20 min to the end of simulation for 060509uv. . 50

Figure 39 Cold pool area of equivalent potential temperature perturbation ($\theta_e'$) < -1 K (top), minimum $\theta_e'$ (middle) and average $\theta_e'$ in the cold pool (bottom) from 20 min to the end of simulation in 060509uv................................................................. 50

Figure 40 Vertical profile of domain averaged rain evaporative cooling (left) and graupel sublimation (dashed) and melting cooling (solid) (right) from 60 to 120 min in 060509uv. .... 51
ABSTRACT

Author: Guo, Mingyang. MS
Institution: Purdue University
Degree Received: August 2018
Title: Aerosol Impacts on Simulated Supercell Thunderstorms in VORTEX2 and VORTEX-SE
Committee Chair: Robin Tanamachi

Environmental factors are found to strongly affect aerosol impacts on convective thunderstorms. In this study, the storm sensitivity to cloud condensation nuclei (CCN) concentration is explored in five aspects: general storm development, hydrometeor, updraft, precipitation and cold pool from a microphysics point of view. Idealized simulations of the 31 March 2016 and 30 April 2016 cases from the Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX)-Southeast (SE) field campaign, and the 5 June 2009 and 9 June 2009 cases from VORTEX2 field campaign are conducted at 6 CCN concentrations (100-3000 cm\(^{-3}\)) with the Advanced Regional Prediction System (ARPS) model. The triple-moment version of the National Severe Storms Laboratory (NSSL) microphysical scheme is used to predict explicitly hydrometeor mass mixing ratio, number concentration and radar reflectivity factor. A budget analysis is performed to understand the impact of each relevant microphysical process on large ice category production, updraft strength, and cold pool intensity. For the simulated supercells, an increase in CCN concentration leads to increasing mass mixing ratios and decreasing sizes of cloud droplets. The opposite is the case for raindrops, as expected, except for the 5 June 2009 case that exhibits a unique reverse trend of cloud mass mixing ratio below 4 km. The cold pool slightly intensifies or weakens with increasing CCN in the VORTEX-SE cases while largely weakens in the VORTEX2 cases. Generally, the influence is more significant in the VORTEX2 than in the VORTEX-SE cases. Other simulated quantities exhibit different responses to CCN enhancement as a function of the environment. For instance,
the mean mass diameter of hailstones peaks at the lowest CCN level in the 30 April 2016 case, at
the highest in the 9 June 2009 case and at an intermediate value in the 31 March 2016 case. Updraft
strength is monotonically increasing with a reduction of CCN concentration in the Jun 5 2009 case
but non-monotonically changing in other cases. The CCN impact on the precipitation rate is also
non-monotonic except for the 5 June 2009 case. The 5 June 2009 case is found to exhibit several
unique features compared with other cases and is further investigated by running sensitivity
simulations using the relative humidity profile, the hodograph or both from the 9 June 2009 case.
CHAPTER 1 INTRODUCTION

Due to the massive damage and loss of life caused by large hail and tornadoes, improving prediction on multiple scales in a region-sensitive manner is necessary and essential for understanding the risk we face and preparing for the future. Supercells are one of the rarest but the most severe storm types in the world, responsible for a disproportionately large fraction of hail and tornado reports (Duda and Gallus, 2010). Nearly all instances of large hail (diameter $\geq 5$ cm), as well as virtually all strong and violent tornadoes (enhanced Fujita scale of EF2 or greater) are associated with supercell storms (Markowski and Richardson, 2011). A typical supercell environment features strong wind shear ($>15$ m s$^{-1}$ over the lowest 6 km, Thompson et al., 2003; Thompson et al., 2007), significant convective available potential energy (CAPE, $>1000$ J kg$^{-1}$, Markowski and Richardson, 2011) and copious amounts of low-level moisture ($>60\%$ below melting level, Thompson et al., 2003; Shimizu et al., 2008); however, these criteria vary across different regions. Supercell environments in the southeastern U.S. (SE) are generally characterized by lower magnitudes of CAPE and deeper tropospheric moisture than those in the Great Plains (GP). These two regions were the focus of the VORTEX2 and VORTEX-SE field campaigns, respectively.

Aerosols in the atmosphere are emitted from both natural (e.g., wildfires, sea foam, desert dust) and anthropogenic (e.g., car engines, industry, fireplaces) sources (Hildemann et al., 1991; Kaufman et al., 2002; Calvo et al., 2013). Aerosols that serve as CCN in deep convection storms can significantly impact the microphysical structure and the behavior of convective storms by altering the cloud droplet size distribution: the so-called second aerosol indirect effect (Warner, 1968; Twomey, 1974). The narrower droplet size distribution is much less collision efficient and thus, significantly modifies the formation processes of rain and other hydrometeors, including ice,
snow, graupel, and hail. In addition, more cloud droplets lifted above the freezing level leads to more latent heat release and thus stronger updrafts aloft in the convection (Andreae et al., 2004; Ramanathan et al., 2001; Wang 2005; Seifert and Beheng, 2006; Van den Heever et al., 2006; Carrió et al., 2010). By impacting the thermodynamic and dynamic processes, aerosols can further show influence on other storm features including precipitation and cold pool.

Moreover, recent simulation studies have shown aerosol effects differ between higher/lower CAPE and moister/drier environments (e.g., Storer et al., 2010; Grant and Van Den Heever, 2014; Kalina et al., 2014; Alizadeh-Choobari and Gharaylou, 2016). Owing to the known sensitivity of severe storms to microphysical differences, studying the impact of aerosols on supercell storms in different environments is of clear societal importance. However, the sensitivity of storm behavior to aerosol effects across different environments are still poorly understood owing to the complex interplay of several factors, and studies have found different or even contradictory results. The factors include but are not restricted to model structural and functional deficiency, observation data assimilation and initial environment features.

The objective of this research is to run simulations on storms in the SE and the GP regions to understand how aerosol impacts on convective thunderstorms change across initial environments characterized by differing CAPE, wind shear and relative humidity. The analysis is mainly from a microphysical point of view by investigating individual microphysical processes involving liquid and ice categories, and the interactions with other storm components.

Chapter 2 reviews the results from the previous literature and presents the hypotheses accordingly. The model configuration and microphysics scheme are described in Chapter 3. The impacts of changing CCN concentration on simulated storm evolution, hydrometeor field, updraft,
precipitation and cold pool are discussed in Chapter 4, and concluding remarks are presented in Chapter 5.
CHAPTER 2 BACKGROUND

Aerosols in the atmosphere play important roles in climate and weather by scattering and absorbing sunlight, participating in chemical reactions and activating cloud droplets (Andreae and Crutzen, 1997; Andreae et al., 2017). The complexity in research on aerosol effects is partly due to the high variability. The particles vary from $10^{-9}$ m to $10^{-5}$ m in diameter and are extremely source-dependent in composition and concentration. The activation of CCN, i.e., water vapor condensation on aerosol particles to form cloud droplets, is affected by the particle size and composition, the type of aerosol, and environmental water vapor supersaturation (McFiggans et al., 2006; Rose et al., 2008; Clavner et al., 2018). The size of an aerosol particle is one of the dominant factors as usually only particles larger than about 30 to 100 nm can activate cloud nucleation (Riipinen et al., 2011). Sulphate aerosols are one of the dominant species that act as CCN and the main source (about 72%) is emission from fossil fuel burning (Charlson et al., 1992; IPCC, 2007; Riipinen et al., 2011), i.e., an anthropogenic source. Organic matters are also found to be crucial in forming cloud but the sources, sinks and the physical and chemical processes involve large uncertainties due to limitation in known species and complexity in composition (Novakov and Penner, 1993; Kanakidou et al., 2005; Fofie et al., 2018).

Debates remain in understanding aerosol influences on several storm features including hydrometeors, updrafts, precipitation, cold pools, and tornadogenesis. Due to the different focuses, a variety of convection modes including mesoscale convective systems, multicellular storms and convective cloud systems except for supercells are concerned in previous research (Table 1). which complicates interpretation and comparison of the results across studies. Compared to environmental factors that have dominant influence on the mode of convective storms including
Table 1 Convection modes simulated in previous studies

<table>
<thead>
<tr>
<th>Single cell</th>
<th>Multicell</th>
<th>Supercell</th>
<th>Mesoscale convective system</th>
<th>Convective cloud and cloud system</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Li et al. (2008)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Van Den Heever et al. (2006)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Van Den Heever and Cotton (2007)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Varble (2018)</td>
</tr>
</tbody>
</table>

CAPE, wind shear (Weisman and Klemp, 1982, 1984; Weisman and Rotunno, 2000) and the moisture profile (Ducrocq et al., 2002; Derbyshire et al., 2004), many researchers have found that overall storm dynamics and mode of convection are less dependent on aerosol concentration (e.g. Lerach and Cotton, 2012; Grant and Van Den Heever, 2014; Loftus and Cotton, 2014; Clavner et al., 2018). The aerosol impacts generally modulate the storm characteristics such as hydrometeor size, precipitation distribution and etc., rather than changing the whole structure, and the overall storm intensity shows minor sensitivity to varying CCN concentration in these studies. Some researchers found that highly polluted environments produce smaller and weaker storms using the double-moment Morrison bulk scheme (Morrison et al., 2009; Morrison and Milbrandt, 2011) in the Weather Research and Forecasting model (Lee, 2011; Fan et al., 2012; Morrison, 2012). Contrarily, Lerach et al. (2008) found that the polluted environment (CCN=2000 cm$^{-3}$) is more conducive to the supercell longevity and tornadogenesis than the clean environment (CCN=600 cm$^{-3}$). Therefore, CCN concentration does show essential impacts on the microphysical processes in the storm and further affects the evolution by interacting with the storm dynamics.
Changes in hydrometeor characteristics with CCN concentration are different across hydrometeor categories. The second aerosol indirect effect (that higher CCN concentration generates more but smaller cloud droplets) has been verified in several studies (e.g., Khain et al., 2005; Lim et al., 2011; Tao et al., 2012). Many studies also found that the impact on rain size is the opposite to that on cloud droplets: more CCN leads to fewer but larger rain droplets (Lim et al., 2011; Loftus and Cotton, 2014; Clavner et al., 2018). However, Lerach et al. (2008) noticed no apparent response in the raindrop size but rather the horizontal distribution relative to the storm updraft location. For larger ice species like graupel and hail, the behaviors associated with supercells are influenced by multiple dynamical, thermodynamical and microphysical processes, including updraft invigoration, accretion of rain, ice and snow, evaporative cooling and others, and such complexity results in the sensitivity to aerosol concentrations being poorly understood. Several studies (Danielson et al., 1972; Khain et al., 2005; Lerach et al., 2008; Kalina et al., 2014) have found that higher CCN concentration leads to larger but fewer hailstones, whereas, Noppel et al. (2010) simulated larger hailstones in the low CCN concentration case. Rosenfeld and Khain (2009) firstly showed a non-monotonic dependence of hail production. Loftus and Cotton (2014) also found that though the mean mass diameter of graupel and hailstones is larger as CCN concentration increases, both mass mixing ratio and number concentration are non-monotonically changing. Carrió et al. (2014) explained the non-monotonic trend shown in some research results by which “the reduction of riming efficiencies due to smaller supercooled cloud droplets increases the fraction of the icephase condensed water mass that is transported to anvil levels as pristine ice crystals instead of being transferred to larger precipitating species.”

In some studies, greater CCN concentration promotes greater updraft speeds (Lynn et al., 2007; Ntelekos et al., 2009; Clavner et al., 2018). On the contrary, the low CCN case produced stronger
updraft in Noppel et al. (2010). Mansell and Ziegler (2014) found no substantial change in maximum vertical wind speed but a noticeable increase of updraft volume as CCN concentration increases.

The response of surface precipitation is found to be highly case-dependent due to the complicated interactions between cloud microphysics and storm dynamics. Noppel et al. (2010) and Lim et al. (2011) found domain-averaged total surface precipitation drops rapidly as CCN concentration increases. Meanwhile, several studies showed a non-monotonic relationship where precipitation increases with CCN concentration rising from clean to polluted environments and then decreases in highly polluted conditions (Li et al., 2008; Kalina et al., 2014; Loftus and Cotton, 2014; Mansell and Ziegler, 2014; Varble, 2018). Furthermore, some demonstrated that changing CCN concentration does not generate a significant change in total surface precipitation but does alter the position of heavy rainfall (Lerach and Cotton, 2012; Clavner et al., 2018). The delayed onset of rainfall as CCN concentration increases was also noticed in several modeling studies (e.g., Tao et al., 2007; Van Den Heever and Cotton, 2007; Storer et al., 2010).

Strongly associated with surface precipitation, the response of cold pool characteristics to CCN concentration is inherently complicated and remains to be fully understood. The intensity and area of cold pools are strongly related to rain evaporation rate, and therefore, apart from precipitation amount, the size of raindrops can also have significant impacts. Also, the presence of larger hydrometeor categories can contribute to cold pool generation by sublimation and melting cooling. Lower CCN environments were found to produce larger and more intense cold pools in several studies (Van Den Heever and Cotton, 2007; Tao et al., 2007; Storer et al., 2010). Kalina et al. (2014), however, found a non-monotonic relationship of cold pool size to increasing CCN, which they explained by noting that less rain evaporation in higher CCN environments was compensated
by more hail melting. Lerach and Cotton (2012) noticed that the coldest region of the cold pool is positioned closer to the main updraft in higher CCN environments since hydrometeors tend to be larger and fall faster, placing heavy precipitation nearer to the updraft core.

The importance of the initial environment quantified by parameters such as relative humidity, CAPE, and vertical wind shear in regulating aerosol impacts has been proposed to explain the disagreements presented in these recent studies. Kalina et al. (2014) pointed out that though raindrop size is assumed to increase with CCN enhancement, in fact in low relative humidity environments, raindrop diameters are about equal in different CCN concentrations due to a higher cloud base preventing cloud droplet collection by rain under 2 km. Grant and Van Den Heever (2014) also performed simulations with a dry layer at different levels and concluded that when the dry layer is positioned near cloud base, the precipitation and the cold pool are intensified at higher CCN concentrations due to more cloud evaporative cooling whereas if the dry layer is higher up, rain evaporation becomes the dominant factor, which results in less precipitation and a warmer cold pool. Similar discussions were also presented in Alizadeh-Choobari and Gharaylou (2016) who found that as CCN increases, rainfall decreases in relatively dry environments while it increases in more humid ones. Lee et al. (2008) found that the sensitivity also differs between environments of high/low CAPE and intense/weak wind shear. With large CAPE (~3000 J kg\(^{-1}\)) and strong wind shear (~25 m s\(^{-1}\) from the surface to 6 km), more aerosols lead to an overall more intense convective storm with a stronger updraft, and cold pool, and more precipitation due to strong interactions between dynamics and microphysics. In contrast, they found that weaker CAPE (500-1500 J kg\(^{-1}\)) environments produce less precipitation and a weaker cold pool at high CCN concentrations due to less cloud condensation and accretion of cloud liquid.
This research is unique from previous studies in two ways. Firstly, it compares the convective storms in the GP and SE regions. Most studies only focused on the GP rather than the SE region which is no less important regarding monitoring severe storm activities. For example, the two regions are comparable in the EF scales of tornados but the majority of the violent ones in the GP happens during the afternoon and early evening hours. In the SE, a considerable percentage of tornadoes take place during the overnight hours, which leaves issuing warnings more difficult and increases dangerous threats to human lives (Gagan et al., 2010). It was also found in some research that the GP region is characterized by more discrete and cluster convective right movers (RM, defined as storms that move right of the mean wind), but that the SE region is characterized by linear RMs, which indicates a higher tornado risk in the latter (Smith et al., 2012; Thompson et al., 2012). Although some performed simulations using idealized soundings with a range of CAPE or humidity values to capture most of the storm-conducive environments, a direct comparison between the two regions has been lacking.

Secondly, this study uses a triple-moment microphysical scheme to explicitly analyze the individual microphysical processes, which relatively few modeling studies have included. Most of the numerical modeling studies used double-moment microphysical schemes and a few used triple-moment schemes for hail only. Kalina et al. (2014) studied hail riming using the double-moment Morrison scheme and found that in dirtiest conditions, riming of rain decreases. However, Loftus and Cotton (2014) augmented a double-moment scheme with a triple-moment hail category and noticed that at higher CCN values, the riming of rain increases as raindrops involved in the hail generation tend to be larger, which leads to bigger hailstones. This disagreement in results may be due to the use of double- or triple-moment schemes. As will be discussed in Chapter 3, triple-moment schemes better parameterize many hydrometeor behaviors and microphysical processes.
Based on the previous studies, we will investigate whether these findings hold true in the very disparate environments of VORTEX2 and VORTEX-SE supercells:

1. (a) As CCN increases, there is more cloud mass but smaller cloud droplets, and less rain mass but larger raindrops. (b) The response of the mass maxing ratio and mean mass diameter of graupel and hail to varying CCN is case-dependent.

2. Updraft strengthens with increased CCN concentration due to more latent heating released from cloud condensation and freezing.

3. Precipitation most likely decreases as CCN increases due to less rain production but may be influenced by the content of graupel and hail.

4. Cold pools weaken at high CCN concentrations because of less rain and inefficient evaporation of larger raindrops.

5. Storms in VORTEX2 environments are more sensitive to CCN changing than in VORTEX-SE environments.
CHAPTER 3 METHODOLOGY

3.1 Model configuration

Soundings from four cases (Figure 1) are used to generate horizontally homogeneous initial environments for the simulations in this study: the 31 March 2016 (033116) and the 30 April 2016 (043016) cases from the VORTEX-SE, and the 5 June 2009 (060509) and the 9 June 2009 (060909) cases from the VORTEX2 field campaign. Observed soundings are used in the simulations of the VORTEX-SE cases. For the 060509 case, a composite sounding of 16 observations (Parker, 2014) is applied since the original observed sounding has a basically neutral boundary layer and thus, generates too much turbulence kinetic energy by the wind shear. The simulation of the 060909 case uses a proximity sounding taken from the Rapid Update Cycle (RUC) weather forecast model so it is smoother than other soundings. The VORTEX-SE cases are characterized by low to moderate mixed-layer convective available potential energy (MLCAPE) and deeper relative humidity (RH) profile while the VORTEX2 cases have a typical high-CAPE environment (more than twice that of the VORTEX-SE cases) but overall lower RH (Table 1). Though there are two dry layers in the 033116 case, one at around 2.5 to 3 km and one at approximately 5 to 6.5 km, compared to the 060509 cases in which the dry layer is at about 1.5 to 3 km, they are positioned at relatively high altitudes, leaving the boundary layer still relatively humid. The 033116 and 060509 cases are tornadic with 0-1 km bulk wind shear (WS) over 80% larger and 0-6 km bulk WS about 25% larger than the non-tornadic ones in the VORTEX-SE and VORTEX2 field campaigns, respectively.
Figure 1 Skew-T plots and hodographs for (a) 033116, (b) 043016, (c) 060509 and (d) 060909, blue dots mark heights at 1000, 3000, 6000 and 12000 m (if applicable).
This study uses version 5.4 of the three-dimensional nonhydrostatic compressible Advanced Regional Prediction System (ARPS) model (Xue et al., 2000, 2001, 2003). The model setup is listed. In the horizontal, the simulated domain has a resolution of 250 m and covers an area of 100 km × 100 km. In the vertical, a stretched grid of 50 levels is applied with a cubic function used for spacing. The average vertical resolution is 400 m, starting from 100 m at the lowest level. The model top is at 20 km AGL. Radiation lateral boundary conditions are used on both east-west and south-north boundaries, and rigid wall on the top and bottom boundaries. The time step is 2 s except for the acoustic waves for which 0.2 s is used. All the cases are run for 2 h with history output every 30 s. Horizontal and vertical advection are both calculated using fourth-order schemes, respectively. Time integration is computed using the leapfrog formulation. Turbulence mixing is calculated with a 1.5-order turbulent kinetic energy scheme. The Coriolis force, radiation physics, and surface physics are neglected for simplicity. Though surface friction has some effects on the propagation of cold pool, it also tends to change the environmental wind profile over time, which is not desired in this study. Dawson et al. (2010) indicated that the effects of microphysics processes appear to overwhelm those by surface fluxes over the relatively short time frames of these simulations and thus lack of surface physics will not have a noticeable influence on the simulation results. Table 2 provides a summary of options used in simulations.

<table>
<thead>
<tr>
<th>Table 2 ARPS options in model simulations</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Horizontal domain size</strong></td>
</tr>
<tr>
<td><strong>Horizontal grid spacing</strong></td>
</tr>
<tr>
<td><strong>Vertical range</strong></td>
</tr>
<tr>
<td><strong>Vertical grid spacing</strong></td>
</tr>
<tr>
<td><strong>Time step</strong></td>
</tr>
<tr>
<td><strong>Duration of model runs</strong></td>
</tr>
<tr>
<td><strong>Turbulence mixing scheme</strong></td>
</tr>
<tr>
<td><strong>Advection scheme</strong></td>
</tr>
<tr>
<td><strong>Time integration scheme</strong></td>
</tr>
</tbody>
</table>
Deep convection is initiated in the model using an updraft nudging technique (Naylor and Gilmore, 2012). The forcing region is 10 km × 10 km horizontally, sharing the same center of the horizontal domain, and 1.5 km vertically, located at 1.5 km AGL. The forcing lasts 900 s at the beginning of the simulations. Each case is initialized with an idealized horizontally homogeneous CCN concentration field at six different levels: 100, 200, 500, 1000, 2000 and 3000 cm⁻³ (denoted as CCN100, CCN200, CCN500, CCN1000, CCN2000 and CCN3000 hereafter). Kalina et al. (2014) found that increasing the CCN concentration beyond 3000 cm⁻³ has a negligible impact on the storm characteristics. Therefore, the values we use should be enough to cover a representative range of aerosol concentrations.

Table 3 Sounding parameters.

<table>
<thead>
<tr>
<th></th>
<th>033116</th>
<th>043016</th>
<th>060509</th>
<th>060909</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface Pressure (hPa)</td>
<td>978.4</td>
<td>989.3</td>
<td>848.3</td>
<td>921.0</td>
</tr>
<tr>
<td>SBCAPE (J kg⁻¹)</td>
<td>1330</td>
<td>814</td>
<td>2346</td>
<td>5882</td>
</tr>
<tr>
<td>MLCAPE (J kg⁻¹)</td>
<td>1040</td>
<td>551</td>
<td>2140</td>
<td>2580</td>
</tr>
<tr>
<td>0-1 km bulk WS (m s⁻¹)</td>
<td>14.16</td>
<td>7.85</td>
<td>7.47</td>
<td>4.12</td>
</tr>
<tr>
<td>0-6 km bulk WS (m s⁻¹)</td>
<td>28.33</td>
<td>23.93</td>
<td>31.36</td>
<td>23.27</td>
</tr>
<tr>
<td>0-2 km mean RH (%)</td>
<td>84.36</td>
<td>76.71</td>
<td>64.71</td>
<td>71.77</td>
</tr>
<tr>
<td>0-6 km mean RH (%)</td>
<td>69.37</td>
<td>76.15</td>
<td>66.89</td>
<td>50.75</td>
</tr>
</tbody>
</table>

3.2 Microphysics Scheme

In bulk schemes, the number spectral density function is often written as a gamma distribution,

\[ n(D) = N_0 D^\alpha e^{-\lambda D}, \]
where D is diameter, $N_0$, $\alpha$ and $\lambda$ are the intercept, shape and slope parameters, respectively (Straka, 2009). In single-moment schemes that predict the third moment–mass–$\lambda$ is diagnosed while $N_0$ and $\alpha$ are constant (e.g., Lin et al., 1983; Cotton et al., 1986). In double-moment ones, the zeroth moment–number concentration is also predicted, meaning both $\lambda$ and $N_0$ are diagnosed independently (e.g., Zeigler, 1985; Murakami, 1990; Reisner, 1998). On top of this, three-moment schemes diagnose all three parameters and predict the sixth moment – reflectivity.

The microphysics scheme used in this study is the triple-moment version of the NSSL bulk microphysics scheme (Mansell, et al., 2010; Mansell and Ziegler, 2013; Dawson et al., 2014), which itself is developed based on an earlier scheme of Ziegler (1985). The scheme predicts six hydrometeor species: cloud droplets, rain, snow, ice, graupel, and hail, all of which are considered as spherical particles. CCN concentration is prognosed; CCN are advected and mixed by the model and are removed from the environment once they are activated by cloud condensation but are not returned as a result of cloud droplet evaporation (Mansell et al., 2013). Two moments of the gamma distribution – the zeroth (number), third (mass) moments are predicted for all the species, and the sixth moment (radar reflectivity) is also predicted for rain, graupel and hail. The triple-moment scheme allows for explicit predictions of microphysical process rates and particle size distributions (Milbrandt and Yau, 2005a, hereafter, MY05a; Mansell et al., 2010; Dawson et al., 2014). The implementation of the sixth moment closely follows the approach in Milbrandt and Yau (2005b, hereafter, MY05b). The scheme allows warm/cold rain processes and dry/wet growth of graupel/hail through a variety of microphysical processes. A full list of source/sink terms for each species is shown in Appendix A. In this section, we only introduce several terms that are of major consideration in the discussion.
Cloud droplets initially form through cloud nucleation when CCN are activated. Cloud can also grow through condensation which happens in supersaturated air. The sum of these two terms is stored in one array named \( q_{wcd} \). The sink terms of the cloud include evaporation, conversion to rain, freezing to cloud ice and accretion by ice, snow, graupel and hail.

Raindrops are initiated from cloud conversion and then grow by collecting cloud droplets. When cold rain and wet growth processes are turned on, graupel/hail shedding and melting will also contribute to total rain mass. The calculation of the size of raindrops produced from shedding and melting is divided based on graupel/hail diameter. For rain melting from graupel/hail particles smaller than 8 mm, the mean diameter of raindrops is based on the size of graupel/hail particles with a maximum of 3 mm. For graupel/hail particles larger than 8 but smaller than 20 mm, mean diameter of rain is adjusted linearly downwards to 3 mm and for particles above 20 mm, mean diameter is set to 3 mm. Shedding occurs when graupel/hail particles grow above 8 mm and the diameter for drops shed from graupel/hail is 1 mm.

Hail initiation is through conversion from graupel over a threshold diameter, which in our simulations is 1 mm. The growth of hailstones represents both dry and wet growth which is separated by Schumann-Ludlam limit (SSL; List, 2010), occurring when a hailstone reaches a surface temperature at which not all the water accreted can freeze and the surplus water is shed. Beyond this threshold, the wet growth mode is activated. Hail can grow by accretion of cloud, ice, snow and rain, which occurs in both wet and dry growth. The process in which ice hydrometeors accrete supercooled cloud droplets or raindrops in the dry growth regime is also called “riming”. In nature, hail mainly grows by accreting liquid water, i.e. raindrops and cloud droplets (MY05b). The number concentration does not change in dry/wet growth. Additionally, hail can grow by direct water vapor deposition.
Graupel comes from either spontaneous rain freezing or three-component freezing that represents rain freezing by contacting with other frozen particles. In our configuration of the scheme, all frozen raindrops are initially categorized as graupel. In terms of dry/wet and other growth processes, graupel behaves nearly the same to hail.

The mean mass diameter of hydrometeors $D_m$ is calculated as in MY05a, according to

$$D_{mx} = \left( \frac{6\rho_a q_x}{\pi \rho_x N_{Tx}} \right)^{1/3},$$

where $\rho_a$ is the air density, $q_x$ is the mass mixing ratio, $\rho_x$ is the bulk hydrometeor density summarized in Table 2, and $N_{Tx}$ is the total number concentration. Subscript $x$ refers to the category of hydrometeor: $c$ is cloud, $r$ is rain, $i$ is ice, $s$ is snow, $g$ is graupel and $h$ is hail. The equation is valid for constant density spheres. According to the description in MY05a, this parameter is useful for diagnosing the size sorting mechanism in the vertical during sedimentation. However, this effect cannot be parameterized in single-moment schemes (MY05a) and in double-moment schemes can be largely overestimated without an adjustment in the distribution shape parameter by diagnostic formula (MY05a) or a correction of the number concentration (Mansell, 2010). Triple-moment schemes allow the shape parameter to increase during sedimentation, narrow the distribution and avoid the unphysical long tails that occur when the shape parameter is constrained to be fixed as in a double-moment scheme without correction (Dawson et al., 2010; Kumjian and Ryzhkov, 2012). Therefore, triple-moment schemes have better performance regarding microphysics parameterization and provide more reasonable simulation results for each hydrometeor category.

<table>
<thead>
<tr>
<th>Category</th>
<th>Cloud water</th>
<th>Cloud ice</th>
<th>Rain</th>
<th>Snow</th>
<th>Graupel</th>
<th>Hail</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bulk density (kg m(^{-3}))</td>
<td>1000</td>
<td>500</td>
<td>1000</td>
<td>100</td>
<td>170-1000</td>
<td>500-1000</td>
</tr>
</tbody>
</table>
3.3 Budget Analysis

A budget analysis of the total mass and energy change in microphysical processes is performed to inspect the role of each process in the behavior of various storm features. The total mass change $\Delta M$ in a period of time is calculated by integrating the mass change in each cell at each time step, that is

$$\Delta M = \sum_{i,j,k,t} (q \rho \Delta x \Delta y \Delta z \Delta t),$$  \hspace{1cm} (3)

where $q$ is the process rate, $\rho$ is air density, $\Delta x$, $\Delta y$ and $\Delta z$ are grid resolutions in the x, y and z directions, and $\Delta t$ is the time step. Based on mass, the energy change $\Delta E$ can be obtained by

$$\Delta E = L \Delta M,$$ \hspace{1cm} (4)

where $L$ is the specific latent heat of vaporization, fusion or sublimation. Further details are found in Dawson et al. (2010).
CHAPTER 4 RESULTS

4.1 CCN Impacts on Storm Evolution

In all the cases, the storms develop from the initial perturbations into a right-moving supercell that remains in the center of the domain and lasts for 2 or more hours, and a left-moving supercell that dissipates or partly moves out of the domain at 120 min. The exception is the 060509 case in which the right-mover nearly dies at 120 min (Figure 2-5). Increasing CCN concentration intensifies the simulated storms in radar reflectivity in the 033116, 043016 and 060909 cases while the opposite is true for the 060509 case. Otherwise, the overall evolution of the storms in terms of reflectivity structure is similar across experiments with differing CCN concentrations. What makes the 060509 case stand out from others is the second pulse in the storm development, i.e., the storm

![Figure 2 Surface radar reflectivity (colored contours) at 60, 90 and 120 min for the 033116 case at CCN100 (a-c) and CCN3000 (d-f).]
intensifies in both radar reflectivity and area after it seems to start weakening at 120 min (Figure 6). For comparison purposes, we will look at the period from 60 to 120 min in each case, which captures the mature stages of each storm and covers the first cycle of the 060509 case. Due to the uniqueness of the 060509 case, we will study the storm evolution and the physical mechanism separately and discuss it later in an individual section.

4.2 CCN Impacts on Hydrometeors

4.2.1 Cloud water

Overall, all cases show an increase of cloud water mass mixing ratio and deduction of cloud droplet mean mass diameter at higher CCN concentrations (Figures 7 and 8). The decrease of cloud mass mixing ratio at about 4.5 km above ground level (AGL) in the 033116 case and the same behavior at approximately 4 km in the 060509 case are likely caused by the presence of a
Figure 4 Same as Figure 2 but for the 060509 case.

Figure 5 Same as Figure 2 but for the 060909 case.
dry layer at that level. The 060909 case also shows a similar trend: cloud water content slowly decreases from 8 to 4 km and then rapidly increases towards the surface due to the relatively moist boundary and overall dry profile above it until 7 km AGL. This influence is more evident in high CCN environments since smaller cloud droplets with higher surface-volume ratio evaporate more efficiently, and thus, higher CCN environments are more sensitive to environmental humidity. Cloud ice is negligible in VORTEX-SE cases where the mass mixing ratio is one to two orders of magnitude smaller than cloud water (not shown) because the updraft is not strong enough (shown later in Section 4.3) to lift the droplets up to freezing while in VORTEX2 cases, the two terms are of the same order.

There are two distinct features in the 060509 cloud mass mixing ratio: the spike at 8 km for CCN over 1000 cm$^{-3}$ and the reverse relationship with CCN concentration as compared with other cases below 4 km. We will discuss the causes of these behaviors later in Section 4.5.
4.2.2 Rain

Rain mass mixing ratio over the whole domain shows a decreasing trend with increasing CCN concentration while the mean mass diameter is increasing in all cases (Figures 9 and 10). On exception is the CCN100 environment in the 033116 case, presenting a slower increase of rain compared to other CCN levels from 4 km downwards. This decrease is likely due to the dry layer, the same reason as the decrease of cloud at that level. The growth of raindrop size accelerates at 3-4 km where graupel stones and hailstones are starting to melt and contribute to large raindrops. In the 060509 case, rain production is quite limited at CCN over 2000 cm\(^{-3}\). The surface mass mixing ratio is about 1/10 of that at CCN100. Again, this feature will be given a further explanation in Section 4.6.

![Figure 7](image_url)

Figure 7 Vertical profile of domain-averaged cloud water mass mixing ratio from 60 to 120 min in (a) 033116, (b) 043016, (c) 060509, (d) 060909.
4.2.3 Graupel and Hail

In contrast to cloud and rain, there are not uniform patterns in the mass mixing ratio and mean mass diameter of graupel and hail and the behaviors are case-dependent (Figures 11-14). Generally, more graupel is associated with smaller graupel particles. The changes of graupel mass and size as a function of CCN concentration are both non-monotonic in the 033116 and 0600909 cases. The peak of mass mixing ratio in the former is at CCN1000, and the peak of mean mass diameter is CCN3000. However, the difference across CCN concentrations is not significant in this case. In the latter, mass peaks at CCN200 to CCN500 and then decreases with CCN while the maximum size is found at CCN3000 and minimum at CCN500. In the 043016 case, the peak of mass is at CCN2000 and generally higher CCN produces more graupel. However, the change of
size shows an opposite trend that it peaks at CCN200 and environments at CCN1000 to CCN3000 produce the smallest graupel stones. The 060509 case presents a monotonically decreasing mass but increasing size of graupel with CCN increasing. For hail, the overall trend is more and larger hailstones at higher CCN (above 2000 cm$^{-3}$) with the exceptions of hail mass mixing ratio decreasing after CCN reaches 500 cm$^{-3}$ in the 060509 case and mean mass diameter peaks at CCN100 in the 043016 case.

Graupel is initiated by rain freezing and grows accreting water or ice particles as introduced in Chapter 3. In these cases, accretion of rain and cloud is the major growth mechanism, two orders larger than accretion of snow and ice (not shown). Due to stronger updrafts that lift raindrops to higher levels in the VORTEX2 cases, more freezing of rain is able to occur, especially for lower CCN environments that produce more rain but smaller raindrops (Figure 15 (g)(j)). Total mass of
Figure 10 Same as Figure 8 but for rain mean mass diameter.

graupel accreting cloud/rain is basically proportional to the amount of cloud/rain available (Figure 15), i.e. more cloud (less rain) at higher CCN concentrations result in more accretion of cloud (less rain). However, accretion of cloud does not peak at CCN3000 in each case, but instead at 1000-2000 cm$^{-3}$ (except for the 060509 case). Carrió et al. (2014) pointed out that when cloud droplet diameters fall below a limit, more fraction will be transported to anvil levels rather than collected by larger ice species since riming is less efficient for smaller droplets. Summing up the three growth processes can explain the behaviors of graupel mass mixing ratio in each case. However, the explanation of trends in mean mass diameter is more complicated. One factor is the size of raindrops that freeze to graupel particles. At the 4-6 km levels where major rain freezing occurs in the VORTEX-SE cases, raindrop mean mass diameter is actually higher at lower CCN, which produces larger embryos. This could be one of the reasons to larger graupel particles at CCN below
Figure 11 Same as Figure 7 but for graupel mass mixing ratio.

500 cm$^{-3}$ than above in the 043016 case: at higher CCN concentrations, the smaller raindrops at the 4-6 km levels lead to smaller graupel stones to begin with while accretion of more cloud droplets cannot compensate for the difference totally. Another factor is (just as in the case of rain), higher CCN concentration leads to a smaller number of graupel stones and therefore each can collect more liquid drops to grow larger. Stronger updrafts that can suspend graupel longer and give it more time to grow, and stronger downdrafts velocity that reduce the time for melting are also factors that impact graupel size but are beyond the scope of this study.

The threshold of graupel conversion to hail in the simulations results in over 60% of the graupel particles converting to hail at all CCN levels and the percentage increases slightly as CCN increases except for the 043016 case (Figure 16). Therefore, total mass conversion from graupel
to hail is mostly determined by total graupel mass (Figure 17 (a)(d)(g)(j)). However, exceptions also happen in the 033116 and the 060909 cases, that graupel mass mixing ratio is lower at CCN3000 than at CCN1000, but more initiates hail because of the higher percentage of graupel stones larger than 1 mm in diameter. Accretion of cloud and rain both increase with increasing CCN concentration (Figure 17) due to more cloud available and more hail initiated. It is not a surprise that growth of hail in the 060509 case is the lowest at high CCN concentrations since the amount of hail initiated is the lowest. The trend of the total hail mass mixing ratio is in agreement with the sum of these individual production processes. Mean mass diameter (Figure 14) follows the same trend as graupel (Figure 12).
Figure 13 Same as Figure 7 but for hail mass mixing ratio.

Figure 14 Same as Figure 8 but for hail mean mass diameter.
Figure 15 Vertical profile of domain-averaged total mass budget in freezing of rain (adgj), graupel accretion of cloud (behk), and graupel accretion of rain (cfil) during 60 to 120 min in 033116 (a-c), 043016 (d-f), 060509 (g-i), and 060909 (j-l).
4.3 CCN Impacts on Updrafts

CCN concentration does not show a universal trend across cases on updraft strength by looking at updraft volumes \((w > 10 \text{ m s}^{-1})\) or maximum vertical wind speeds (Figure 18) especially after 60 min when the storm reaches its mature stage. For the 033116 case, the trend of updraft volume is non-monotonic and peaks at CCN1000, though the difference among environments with CCN over 200 cm\(^{-3}\) is not significant. For the 043016 case, it can be grouped by CCN over 500 cm\(^{-3}\) with an average volume (between 60 and 120 min) of about 35 km\(^3\) and below with roughly 11 km\(^3\). The same trend is not shown in maximum vertical velocity. A dramatic difference shows in the 060509 case, that updraft volume decreases with CCN enhancement especially for CCN over 2000 cm\(^{-3}\). The average volume between 60 and 120 min at CCN over 2000 cm\(^{-3}\) is about 1/4 of that below. For the 060909 case, there is no apparent relationship between CCN concentration and updraft volume or maximum vertical wind speed. The volumes at all CCN levels shrink to about half the size after 80 min because the left movers are moving out of the domain. To test the previous theory in Khain et al. (2005) that due to low collision efficiency and thus delayed formation of rain, the duration of the diffusion droplet growth increases, adding more latent heating from cloud
condensation, we looked into the amount of latent heat released by cloud condensation (Figure 19), which is also the largest heating term compared to other microphysical processes (not shown).
Figure 18 Updraft volume of $w>10 \text{ m s}^{-1}$ (top) and maximum vertical velocity (bottom) over the whole simulation time period in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

Some common features can be observed between cloud condensation latent heating and updraft strength. For example, the least latent heating at CCN $100 \text{ cm}^{-3}$ matches with the weakest updraft in the 033116 case. Also, the considerably small updraft volume is related to the lowest latent heat release at CCN over $2000 \text{ cm}^{-3}$ in the 060509 case. However, the relationship is not monotonic, and the microphysics process alone can hardly account for the full behaviors of updrafts. Where
the results of this study also show differences from the Khain et al. (2005) paper is that the 060509 case demonstrates an opposite trend in cloud condensation changing with CCN concentration.

Figure 19 Total latent heating released by cloud condensation and freezing over the whole simulation time period in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.

4.4 CCN Impacts on Precipitation

Increased CCN concentration delays the formation of precipitation; the maximum rainfall rate is delayed by 10 to 60 minutes in different cases, shorter in the VORTEX-SE cases but longer in the VORTEX2 cases (Figure 20). For the values of precipitation rates, 060509 case again presents a dramatic gap across the CCN range, which is expected due to a similar trend in the mass mixing ratios of precipitation species (Figures 9, 11 and 13). However, the impact on other cases is not as obvious. Generally, the precipitation rate peaks at CCN200 to CCN500 and the highest 60-120 min average is less than 30% higher than the lowest. Because of the distinct features and the very limited precipitation at CCN over 2000 cm$^{-3}$ in the 060509 case, we will leave it to Section 4.6 and only discuss the other cases for now.
Though the impact on total precipitation rate is not evident, CCN concentration alters the horizontal pattern of precipitation. The heaviest precipitation region tends to locate closer to the rear flank downdraft (RFD) in higher-CCN environments while it is closer to the forward flank downdraft (FFD) in lower-CCN environments (Figures 21, 22). This is due to the size distribution of precipitation particles, with larger raindrops, graupel stones and hailstones at higher CCN concentrations placed nearer to the updraft core (Figures 23-28). Larger particles that fall out of the main updraft tend to land horizontally nearer to the updraft due to higher terminal velocity. Though the size of graupel and hail in the 043016 case is the largest at CCN below 200 cm$^{-3}$, this overall feature is still valid since the mass mixing ratio of large ice particles is relatively low (Figures 11,13) and thus rain becomes the dominant precipitation species.

Figure 20 Total surface precipitation rate from 20 min to the end of simulation in (a)033116, (b)043016, (c)060509 and (d)060909.
4.5 CCN Impacts on Cold Pools

The change of the size and the intensity of the cold pool (equivalent potential temperature perturbation $\theta_e' < -1$ K) as CCN concentration decreases is different between the VORTEX2 and VORTEX-SE cases (Figure 25). Overall the cold pool characteristics are more sensitive to CCN changing in the VORTEX2 cases than in the VORTEX-SE cases (Figure 26), which agrees with Carrió et al. (2014) that environments with drier low-level layers are more affected by differences in CCN concentrations. For average from 60 to 120 min, from CCN 100 cm$^{-3}$ to 3000 cm$^{-3}$, the percentage changes of each quantity are listed in Table 4.

This different sensitivity can be explained by looking into the major latent cooling processes. Due to less but bigger raindrops under higher CCN environments, latent cooling by rain evaporation near the surface is reduced in all cases (Figure 27). However, this loss of evaporative cooling can be compensated for by other microphysical processes. Graupel melting and sublimation are two dominant processes causing latent cooling above the surface. The cold air produced can descend and add to the formation of cold pools. In the VORTEX-SE cases, cooling by graupel processes is higher in more polluted environments (Figure 28) and thus can, to some extent, compensates for the lack of rain evaporation. However, in the VORTEX2 environments, both graupel melting and sublimation are decreasing with CCN increasing in both cases. Therefore, there is no compensation effect and the overall reduced latent cooling leads to weaker and smaller cold pools. The contribution of graupel processes to cold pool construction also differs between the two field campaign environments. In the VORTEX-SE cases, rain evaporation is the dominant cooling source while graupel sublimation and melting together take up about 20-70% of the rain evaporative cooling (Table 5). On the contrary, the graupel process cooling in the VORTEX2 cases
Figure 21 Surface precipitation rate (colored contours) at 60 min at CCN 100 cm\(^{-3}\) (a-c) and 3000 cm\(^{-3}\) (def), vertical velocity (line contours) at 3 km with levels of 5 m s\(^{-1}\) (thin solid) and 15 m s\(^{-1}\) (thick solid) and downdraft of -3 m s\(^{-1}\) (dashed) in 033116 (ad), 043016 (be), and 060909 (cf).

Figure 22 Same as Figure 21, but at 90 min.
Figure 23 Surface mean mass diameter of rain (ad), graupel (be) and hail (cf) at 60 min at CCN 100 cm\(^{-3}\) (a-c) and 3000 cm\(^{-3}\) (d-f), vertical velocity (line contours) at 3 km with levels of 5 m s\(^{-1}\) (thin solid) and 15 m s\(^{-1}\) (thick solid) and downdraft of -3 m s\(^{-1}\) (dashed) for 033116.

Figure 24 Same as Figure 23, but at 90 min.
Figure 25 Surface mean mass diameter of (a) rain, (b) graupel and (c) hail at 60 min at CCN 100 cm\(^{-3}\) (top) and 3000 cm\(^{-3}\) (bottom), vertical velocity (line contours) at 3 km with levels of 5 m s\(^{-1}\) (thin solid) and 15 m s\(^{-1}\) (thick solid) and downdraft of -3 m s\(^{-1}\) (dashed) for 0430 16.

Figure 26 Same as Figure 25, but at 90 min.
Figure 27 Cold pool area of equivalent potential temperature perturbation (θe') < -1 K (top), minimum θe' (middle) and average θe' in the cold pool (bottom) from 20 min to the end of simulation in (a) 033116, (b) 043016, (c) 060509 and (d) 060909.
Table 5 Percentage change of cold pool parameters

<table>
<thead>
<tr>
<th>From CCN3000 to CCN100</th>
<th>Area</th>
<th>Min $\theta'_e$</th>
<th>Mean $\theta'_e$</th>
</tr>
</thead>
<tbody>
<tr>
<td>033116</td>
<td>-22.3%</td>
<td>25.0%</td>
<td>19.0%</td>
</tr>
<tr>
<td>043016</td>
<td>30.6%</td>
<td>0.8%</td>
<td>2.2%</td>
</tr>
<tr>
<td>060509</td>
<td>180.4%</td>
<td>-96.3%</td>
<td>-404.5%</td>
</tr>
<tr>
<td>060909</td>
<td>17.1%</td>
<td>-6.3%</td>
<td>-80%</td>
</tr>
</tbody>
</table>

Figure 28 Surface equivalent potential temperature (colored contours) at 60 min at CCN 100 cm$^{-3}$ and 3000 cm$^{-3}$, and 90 min at CCN 100 cm$^{-3}$ and 3000 cm$^{-3}$, vertical velocity (line contours) at 3 km with levels of 5 m s$^{-1}$ (thin solid) and 15 m s$^{-1}$ (thick solid) and downdraft of -3 m s$^{-1}$ (dashed) in 033116 (a-d), 043016 (e-h), 060509 (i-l), and 060909 (m-p).
is much more enhanced and together exceeds the contribution of rain evaporation. Especially in the 060509 case, both melting and sublimation processes take place nearer to the surface than in other cases, which strengthen the influence. All these factors add up to the much weaker cold pool in the higher-CCN case for 060509 which is ultimately due to a much weaker storm.

4.6 The 060509 Case

The 060509 case presents several distinct features from other cases, and therefore, it will be discussed separately in this section.

Firstly, the profile of cloud mass mixing ratio shows a spike compared to the 060909 case at 8 km for CCN over 1000 cm\(^{-3}\) as mentioned in 4.2 (Figure 7(c)). The spike does not exist anymore,

![Figure 29 Vertical profile of domain averaged rain evaporative cooling from 60 to 120 min in (a) 033116, (b) 043016, (c) 060509, and (d) 060909.](image-url)
and the whole profile resembles the 060909 case if both cloud water and ice are considered instead of only the former (Figure 29(a)(b)). One explanation is that the 060909 case has a slightly warmer sounding with averaged temperature about 5°C higher than the 060509 case (Figure 29(c)), so the freezing level is pushed higher.

Secondly, the cloud mass mixing ratio shows a reverse relationship with CCN concentration (decrease with CCN enhancement) below 4 km. This reduction of the cloud is accompanied by decreased production of all the precipitation categories including rain, graupel, and hail. Apart from hydrometeors, according to the analysis of the updraft, precipitation, and cold pool, apparently, though the CAPE is large enough to support intense storms and wind shear is also

Figure 30 Same as Figure 27 but for graupel sublimation (dashed) and graupel melting (solid).
Table 6 Ratio of total graupel sublimation and melting cooling, and rain evaporative cooling.

<table>
<thead>
<tr>
<th></th>
<th>033116</th>
<th>043016</th>
<th>060509</th>
<th>060909</th>
</tr>
</thead>
<tbody>
<tr>
<td>CCN100</td>
<td>0.24</td>
<td>0.18</td>
<td>2.31</td>
<td>1.14</td>
</tr>
<tr>
<td>CCN3000</td>
<td>0.49</td>
<td>0.71</td>
<td>1.80</td>
<td>1.42</td>
</tr>
</tbody>
</table>

supportive of tornadic supercells, the components of the 060509 sounding do not support a storm well enough for its growth and intensification under highly polluted conditions, in this case, CCN over 2000 cm$^{-3}$.

To understand why even with similar CAPE to the 060909 case, the 060509 sounding fails to support storms under highly polluted conditions, we tested the influence of environmental relative humidity and horizontal wind profiles separately by conducting a set of factor separation experiments based on Stein and Alpert (1992), to replace 1) the relative humidity profile, 2) the wind field, and 3) both in the 060509 case with those in the 060909 case, respectively (Figure 30).

Figure 31 Vertical profile of domain averaged cloud water and ice mass mixing ratio from 60 to 120 min in (a) 060509 and (b) 060909, and (c) sounding temperature profiles.

The profiles are taken by interpolating the AGL heights in the 060509 case onto the corresponding values in the 060909 case. Also, relative humidity interpolation starts from 1 km so that the boundary layer, and thus the SBCAPE and MLCAPE, is unaffected (since CAPE is a strong function of boundary layer RH). The three factor-separation simulations are referred to as 060509rh, 060509uv, and 060509rhuv, respectively. With RH profile changed, the storm
Figure 32 As in Figure 1 but for the 060509 case after interpolation onto the 060909 case vertical levels.

Figure 33 Surface radar reflectivity (colored contours) at CCN 100 cm$^{-3}$ at 60 min (a-c), 90 min (d-f) and 120 min (g-i) for the 060509rh (adg), 060509uv (beh), and 060509rhuv (cfi).
significantly weakens and quickly dissipates at CCN100 while there is hardly a storm at CCN3000 (Figure 31(b), 32(b)) because of the reduction of moisture from 1 km to 7 km. The 060509uv case presents an intensified and more organized storm especially at CCN3000 (Figure 31(c)(d), 32(c)(d)). Also, the storm is more evenly separated because the more straight-line like hodograph in the 060909 case has a tendency to produce equally developing left and right movers while the more curved hodograph in the 060509 case favors a prevalent right mover (Weisman and Rotunno, 2000). Another difference is that there is no longer a second pulse in the storm evolution (not

Figure 34 Same as Figure 35 but for CCN3000.
shown). The 060509rhuv case resembles the 060509uv case. According to the results, it is likely that the wind profile is one of the key factors leading to the unusually weak storms at CCN over 2000 cm\(^{-3}\). Therefore, we will focus on the comparison between the 060509 and the 060509uv cases.

The mass mixing ratios of all hydrometeors considered are enhanced in the 060509uv case at CCN2000 and CCN3000 while the difference is not evident at lower concentrations (Figure 33). Cloud water at 6 km AGL has increased by 50% and at 3 km AGL by 30%. The maximum rain mixing ratio is 3 times larger, the graupel about 5 times larger and the hail more than 6 times larger.

![Figure 35 Vertical profile of domain averaged mass mixing ratio from 60 to 120 min of (a) cloud water, (b) rain, (c) graupel and (d) hail for 060509uv.](image)

This increase of cloud water is the essential factor since it is the ultimate water source for the other larger hydrometeor categories, directly or indirectly. In order to understand the cloud behavior, a new term cloud evaporation fraction is defined as the ratio between the mass mixing ratio of
Figure 36 Vertical profile of domain averaged evaporation fraction from 60 to 120 min in (a)033116, (b)043016, (c)060509, (d)060909, (e)060509uv, and (f)060509rhuv

evaporated cloud and total cloud water at a certain level. Thus, this ratio represents how much cloud water evaporates compared to the rest that comes from several processes (see APPENDIX) and remains at that level. In other cases, except for the 060509 case, evaporation efficiencies at CCN over 2000 cm\(^{-3}\) only exceed that at lower concentrations below around 3 km whereas, it starts at 6 km in the 060509 case (Figure 34). This unusually high cloud evaporation fraction in the
060509 case is thought to result from the combination of the low low-level relative humidity and high wind shear, which results in more entrainment than other cases. The smaller cloud droplets in higher CCN concentrations evaporate more readily especially when the environment is relatively dry. Entrainment brought by strong wind shear can be detrimental to the cloud formation process in this situation and thus, the storm would struggle to develop. The wind shear in the 060509 case is much larger than that in the 060909 case (0-1 km shear 45% more and 0-6 km shear 25% more). Consequently, the depletion of cloud leads to less rain production by collision and consequently less larger ice categories since the accretion of cloud water and rain is the major growth mechanism. The 060909 wind profile has smaller wind shear and thus does not harm cloud initiation as much. Therefore, the cloud evaporation fraction in the 060509uv and 060509rhuv cases no longer shows a deep layer of high values at CCN2000 and CCN3000 (Figure 34 (e)(f)).
Figure 38 Total surface precipitation rate from 20 min to the end of simulation for 060509uv.

Figure 39 Cold pool area of equivalent potential temperature perturbation ($\theta_e'$) < -1 K (top), minimum $\theta_e'$ (middle) and average $\theta_e'$ in the cold pool (bottom) from 20 min to the end of simulation in 060509uv.
Apart from the differences in hydrometeors, the updraft in the 060509uv case also shows enhancement at CCN over 1000 cm$^{-3}$ (Figure 35). The average updraft volume from 60 to 120 min at CCN1000 is about 50% larger than in the 060509 case, CCN2000 230% and CCN3000 200%; the maximum vertical velocity is also 10% higher at CCN2000 and CCN3000. The gap between the updraft strength at higher and lower CCN concentrations becomes less dramatic.

As the overall storms in the more highly polluted environments intensify, the 060509uv case also shows an increase in precipitation amount and cold pool strength compared to the 060509 case (Figures 36, 37). However, the gap between CCN below 200 cm$^{-3}$ and CCN over 2000 cm$^{-3}$ is still quite significant because both rain evaporation cooling and graupel sublimation and melting cooling are still rapidly decreasing with CCN enhancement (Figure 38).

Figure 40 Vertical profile of domain averaged rain evaporative cooling (left) and graupel sublimation (dashed) and melting cooling (solid) (right) from 60 to 120 min in 060509uv.
CHAPTER 5 CONCLUSION AND DISCUSSION

This research investigates the sensitivity of several storm features--including hydrometeor fields, updraft volume and magnitude, precipitation rates, and cold pool area and strength--to changes in the initial assumed environmental CCN concentrations in three-dimensional simulations of four supercell thunderstorms. Two occurred during the VORTEX-SE and two during the VORTEX2 field campaigns. The NSSL triple-moment microphysical scheme is employed to predict hydrometeor sizes and microphysical process rates explicitly. The results are examined across 6 CCN values from 100 cm\(^{-3}\) to 3000 cm\(^{-3}\).

Cloud content increases but droplet size decreases in the VORTEX-SE cases and the 060909 case, which is as expected from our hypotheses. However, a reduction of cloud mass mixing ratio with the increase of CCN is found in the 060509 CCN2000 and CCN3000 cases below 4 km, which differs from previous findings and demands further investigation. The formation of cloud in environments with relatively dry low-level layers above the boundary layer can be vulnerable to wind shear due to environmental entrainment evaporating the smaller cloud droplets at higher CCN concentrations. Rain mass mixing ratio decreases while mean mass diameter increases as CCN increases, which is consistent with many previous studies expect those simulating cases in relatively dry environments.

Through a budget analysis of microphysical process mass production, we demonstrated that the mass mixing ratio of graupel and hail is affected by the content of the initiating category (rain and graupel, respectively) and the availability of the riming categories, which in this case, cloud water and rain. We also showed that the mean mass diameter of large ice species is decided by the number and the size of the initiating species, availability of the riming species and storm dynamics. Both quantities are highly case-dependent and exhibit a mostly non-monotonic relationship to CCN.
concentration, which is similar to some previous studies (Rosenfeld and Khain, 2009; Carrió et al., 2014).

Updraft strength and surface precipitation rate are both non-monotonically changing with CCN enhancement except for the rapid drop in the 060509 case that has relatively dry low-level layers and strong wind shear. Instead, high CCN concentration affects precipitation by delaying the onset of rainfall and altering the distribution of heavy precipitation, which is consistent with the conclusions in several studies (Tao et al., 2007; Van Den Heever and Cotton, 2007; Storer et al., 2010).

In all the cases except 033116, the cold pool shrinks and weakens as CCN concentration decreases whereas, in the 033116 case, it enlarges and slightly intensifies. The CCN influence is more significant in the VORTEX2 than in the VORTEX-SE cases. This is explained through a budget analysis of latent cooling in microphysical processes. The combination of the loss of rain evaporative cooling with the gain (loss) of graupel sublimation and melting cooling is the reason to smaller (bigger) differences of cold pool sizes and strengths across CCN concentrations. Similar results were found in Kalina et al. (2014) in which only hail instead of both graupel and hail categories were included and increasing hail melting was found to compensate the decreasing rain evaporation as CCN concentration increases.

We also analyzed the cause of the unusually weak storms in the 060509 CCN2000 and CCN3000 conditions by replacing the relative humidity profile and hodograph with those in the 060509 case, respectively. We found that the dry layer at low levels and strong wind shear may be one explanation for the results. Entrainment brought by the strong wind shear evaporates the smaller cloud droplets in higher CCN environments and thus, harm the onset of convective storms especially those with relatively dry low-level layers.
In fact, most studies that found monotonic relationships in graupel/hail production and updraft, and precipitation characteristics with CCN concentration examined three or fewer different CCN concentrations, which is not enough to capture the changes across the whole CCN variation. It has been demonstrated in several previous studies as mentioned in Chapter 2 and is expected for this research that these quantities show a non-monotonic trend related to CCN concentration change: increasing (decreasing) to a certain tipping point around 500 cm\(^{-3}\) to 1000 cm\(^{-3}\) while then decreasing (increasing) above this value. This indicates that the presence of aerosols has complex effects on many convective storm characteristics. The initial increase of CCN concentration contributes to more cloud droplets and thus enhancing the microphysical processes. However, further increasing the concentration will lead to much smaller cloud droplets that have low collision and riming efficiencies, thus delaying the formation of larger precipitation species. Furthermore, the inefficiency results in more cloud droplets being lofted higher by updrafts, often transported to the anvil instead of participating in the generation of larger hydrometeors.

Overall, the simulated supercell thunderstorms are more sensitive to CCN concentration changing in environments with lower low-level RH that are typical of the GP environments compared to the SE region. Carrió et al. (2014) also reached a similar conclusion from the simulation of a single-cell hailstorm. This indicates that microphysical schemes that better parameterize aerosol-cloud interaction are necessary for modeling and predicting deep convective storms in the GP region.

Though we did not discuss tornadogenesis in this study, it is an important topic in understanding behaviors of supercells which are responsible for a disproportionally large percentage of tornadoes as introduced in Chapter 1. According to some previous research (Markowski et al., 2002; Lerach et al., 2008; Snook and Xue, 2008) environments with weaker cold pools are more favorable for
tornadogenesis while Lerach and Cotton (2012) came to an opposite conclusion in a lower low-level moisture case (20% less water vapor mixing ratio below 800 mb than in Lerach et al., 2008). This indicates a difference in the CCN impacts on tornadogenesis potential between the GP and SE environments, which requires further research, such as simulations with higher horizontal resolutions to capture the tornadic wind field characteristics.

In many of the simulations in this study, part of the left-moving supercell moved out of the domain towards the end of the simulation period. For future work, the simulation domain is to be enlarged in order to capture the entire storm up until 120 min.

The approach in this research to directly take from the 060909 sounding and hodograph is an initial step to test the storm sensitivity to environmental factors based on the unique features of the 060509 case. To fully understand the mechanism of such behaviors, a set of sensitivity tests by stepwise modifying the 060509 relative humidity profile and wind field is proposed for future work.

Finally, we will also investigate more deeply the physical explanations on the non-monotonic behaviors in storm characteristics.
APPENDIX

Microphysical Terms of Mass Production

The left-hand side term is total mass change rate, denoted by $q_{x.y}$, where $q$ means rate, $x$ is the hydrometeor species ($w$ for cloud water, $i$ for cloud ice, $r$ for rain, $s$ for snow, $h$ for graupel and $hl$ for hail), $y$ stands for increase ($i$) or decrease ($d$) of the mass. The right-hand side terms are microphysical process rates, denoted by $q_xA(x)$, where $A$ represents the process ($cnd$ for condensation, $frz$ for freezing, $cev$ for evaporation, $dpv$ for deposition, $sbv$ for sublimation, $mlr$ for melting, $shr$ for shedding, $ac$ for accretion and $cn$ for conversion), and if the process involves more than one category, a last character $x$ is added to specify the affected category.

For snow accreting rain, the affected categories depend on the mass mixing ratio of the two species. If there is small amount of both snow and rain ($<10^{-4}$ g g$^{-1}$), then the process only increases the mass mixing ratio of snow. Otherwise, the increased mass goes to graupel. A coefficient $i_{l2}$ is multiplied to relevant terms. Some processes occur only when temperature is above 0 while some only when temperature is below 0. A coefficient $i_{l5}$ is used to control the relevant processes. The coefficients are defined as follows:

$$i_{l2} = \begin{cases} 1, & qs < 10^{-4} \text{ and } qr < 10^{-4} \\ 0, & \text{otherwise} \end{cases}$$

$$i_{l5} = \begin{cases} 1, & T < 0 \\ 0, & \text{otherwise} \end{cases}$$

We only list the terms of mass production here. Particle number concentration is calculated in a similar fashion with some differences, which is not the focus of this research. The change of mass mixing ratio and microphysical process rates are as follows:
\[ qwi = qwcn 
\]
\[ qwd = il5(qwfrz + qiacw) + qwcev + qwcnr + qsacw + qracw + qhacw + qhlacw \]  
(A3)
\[ qii = il5(qwfrz + qidpv + qiacw) \]  
(A4)
\[ qid = il5(qisbv + qscni + qsaci + qraci) + (1 - il5)qimlr + qhcni + qhaci + qhlaci \]  
(A5)
\[ qri = qracw + qrcnw + (1 - il5)(qimlr + qsmlr + qhmlr) + qhlmlr + qssh + qhshr + qhlshr \]  
(A6)
\[ qrd = il5(qrfrz + qiacr) + qrcev + qscar + qhacr + qhlacr \]  
(A7)
\[ qsi = il5(qsdpv + qscni + qsaci + qwfrz) + il2(qsacr) + qsacw \]  
(A8)
\[ qsd = (1 - il5)qsmlr + qssh + qssbv + (1 - il2)qrcs + qhacs + qhlacs \]  
(A9)
\[ qhi = il5(qrfrz + qhdpv) + (1 - il2)(qrcs + qscar) + qhacr + qhaci + qhacw + qhacs + qhcni + qhcns \]  
(A10)
\[ qhd = (1 - il5)qhmlr + qhshr + qhsbv \]  
(A11)
\[ qhli = qhlcnh + il5(qhldpv) + qlacw + qhlacr + qhlaci + qhlacs \]  
(A12)
\[ qhld = (1 - il5)qhlmlr + qhlshr + qhlsbv \]  
(A13)
REFERENCES


