Investigating the effects of ice-forming Saharan dust aerosols on tropical deep convection using spectral bin microphysics

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Investigating the Effects of Ice-forming Saharan Dust Aerosols on Tropical Deep Convection Using Spectral Bin Microphysics

by

Matthew S. Gibbons

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Investigating the Effects of Ice-forming Saharan Dust Aerosols on Tropical Deep Convection Using Spectral Bin Microphysics

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Abstract

Aerosol effects on cloud and precipitation formation remain a significant source of uncertainty in the study of weather and climate. Aerosols can impact cloud and precipitation formation by functioning as cloud condensation nuclei (CCN), giant cloud condensation nuclei (GCCN) and/or ice nuclei (IN) affecting subsequent cloud microphysical processes. Aerosol effects on clouds are tightly interconnected with cloud dynamic and thermodynamic variables, some of which are currently impossible or infeasible to observe with existing sensors. Numerical models can be used to untangle aerosols effects from cloud dynamics and thermodynamics, but model results can be affected by the complexity of the parameterizations used to represent cloud microphysical processes in the model. Deep convective clouds (DCC) are important sources of precipitation and play a strong role in both regional and global circulation, with tropical convection being particularly significant. However, understanding of ice processes within these clouds is still limited due to the dynamically and thermodynamically complexities of DCCs and the lack of parameterizations that directly connect ice formation with aerosols. Therefore, to better understand the impacts of dust aerosols on DCC systems reported by previous observational studies, a case study in the tropical eastern Atlantic was investigated using the Weather Research and Forecasting (WRF) model coupled with a Spectral Bin Microphysics (SBM) model. A detailed set of ice nucleation parameterizations linking ice formation with aerosol particles have been implemented in the SBM.

Increasing ice nuclei (IN) concentration in the dust cases results in the formation of more numerous small ice particles in the heterogeneous nucleation regime (between -5°C and -38°C) compared to the background (Clean) case. Convective updrafts are invigorated by increased
latent heat release due to depositional growth and riming of these more numerous particles, which results in increased overshooting and higher convective core top heights. Homogeneous ice formation is reduced in the dust cases as IN concentration is increased, due to more liquid drops converting to ice by freezing or riming before reaching -38°C and reduced peak supersaturation values from increased diffusional growth. Local IN activation in the stratiform regime contributes to increased cloudiness in the heterogeneous nucleation regime.

Competition between the more numerous particles for available water vapor during diffusional growth and available smaller crystals/drops during collection reduces particle growth rates and shifts precipitation formation to higher altitudes in the heterogeneous nucleation regime. A greater number of large snow particles form in the dust cases, which are transported from the core into the stratiform regime and sediment out quickly. Together with reduced homogeneous ice formation, fewer particles form within and/or are transported into the anvil regime. This shifts the stratiform/anvil cloud occurrence frequency to warmer temperatures and reduces anvil cloud extents.

Total surface precipitation accumulation is reduced proportionally as IN concentration is increased, due to less efficient graupel formation reducing convective rain rates. Stratiform precipitation accumulation is increased due to greater snow formation and growth, but does not counteract the reduced convective accumulation. Rimming efficiency in the dust cases is reduced due to smaller cloud ice crystals, resulting in smaller graupel sizes overall. Ice particle aggregation occurs earlier in the simulation and over a wider temperature range in the dust cases, which increases snow formation in the heterogeneous nucleation regime. Radar reflectivity values are increased in the dust cases at temperatures below 0°C in both the convective and stratiform regimes due to more large snow particles. More numerous small ice/graupel particles
that form in the heterogeneous nucleation regime in the dust cases melt and reduce reflectivity values in the convective core near the surface, consistent with case study observations.
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Chapter 1

Introduction

1.1 Aerosol -Cloud-Precipitation Interactions (ACPI)

Aerosol effects on cloud and precipitation formation remain a significant source of uncertainty in the study of weather and climate. Aerosols may directly impact the atmosphere by absorbing/scattering radiation in the atmosphere and affecting relative humidity and atmospheric stability (Wang, 2013). Aerosols may also indirectly impact cloud and precipitation formation by functioning as cloud condensation nuclei (CCN), giant cloud condensation nuclei (GCCN) and/or ice nuclei (IN) affecting subsequent cloud microphysical processes. This is typically referred to as the “aerosol indirect effect” (AIE). Aerosol effects on clouds are tightly interconnected with cloud dynamic and thermodynamic variables, some of which are currently impossible or infeasible to observe with existing sensors. Numerical models can be used to untangle aerosols effects from cloud dynamics and thermodynamics, but model results can be affected by the complexity of the parameterizations used to represent cloud microphysical processes in the model (Khain et al., 2015; Fan et al., 2016b). Therefore, observational and modeling studies of ACPI have shown different and occasionally conflicting results relating to the effect of aerosols on cloud and precipitation processes. These studies report that aerosol may either enhance or suppress convective intensity and/or precipitation formation depending on the type of cloud being affected, the type of aerosol involved, and the background environmental conditions (Khain and Pokrovsky, 2004 Khain et al., 2004, 2005, 2008; van den Heever et al.,
Deliberate attempts to enhance precipitation formation in various types of clouds by “seeding” them with salt particles, silver iodide, dry ice, etc. have been studied since the mid-1940s. While some indications exist that cloud seeding may potentially enhance precipitation formation, the scientific credibility of many of these studies is in question (Bruintjes, 1999). The effects of cloud seeding may be insignificant compared to the natural variability of precipitation and accurately tracking the specific cloud volume affected by seeding can be a substantial challenge as well (Rosenfeld, 2007; Pokharel et al., 2015). In many studies of ACPI, clouds forming in elevated aerosol environments frequently exhibit reduced cloud drop effective radii as a result of a greater number of smaller drops forming from aerosols activating as CCN (Andreae et al., 2004; Koren et al., 2005). This can result in less efficient collision-coalescence processes (Khain et al., 2005) which shifts the formation of precipitation to higher altitudes in the clouds. Condensation and evaporation processes are affected by the altered drop size distribution and number concentration, resulting in changes to the location and intensity of latent heat release within the cloud (Khain et al., 2005; Rosenfeld et al., 2008). The higher droplet concentrations induce greater condensation and latent heat release, resulting in stronger convective updrafts and the formation of taller and wider clouds (Frederick, 2006; Zhang et al., 2007). Increased evaporation of smaller rain drops can potentially result in stronger cold pool formation and enhanced secondary convection (Khain 2009, Lee et al., 2010). Conversely, other studies have noted that the formation of larger rain drops due to enhanced rain drop collision-coalescence limits evaporation and weakens the cold pool (Altaratz et al. 2007; Berg et al. 2008; Lerach et
Aerosol indirect effect related changes to cloud macrophysical properties are frequently attributed solely to thermodynamic convective invigoration due to increased latent heat release resulting from changes liquid and/or ice particle number concentrations, greater total condensate mass and subsequent changes to diffusional growth processes. However, modeling studies have suggested that thermodynamic invigoration can be insignificant or even suppressed for clouds with a cold base or for clouds developing in a dry and/or high wind shear environment (Fan et al., 2009, 2012b, 2013; Li et al., 2008b; Khain et al., 2005, 2008a; Tao et al., 2007; Lebo et al., 2012; Lebo and Seinfeld, 2011). For example, the study by Fan et al. (2013) involving simulations of DCC in three different regions, suggested that the observed taller and wider clouds could be better explained by changes to microphysical properties such as the particle size distribution. Thermodynamic invigoration by increased latent heat release did not unanimously occur in the study when polluted conditions were simulated, although increased cloud fraction and cloud top height were present. The study noted that the reduced hydrometeor sizes in the polluted case allowed greater cloud mass to be detrained from the convective core, and decreased particle fallout speed that slows down the cloud anvil dissipation.

Outside of aerosol number concentration and particle chemistry, ACPI are sensitive to the background environment in which the clouds are forming. Factors such as relative humidity (RH), CAPE, and vertical wind shear are particularly important. The effect of aerosols on drop size distribution and precipitation can be negligible in a dry environment (40% surface RH) and become more significant in humid environments (60-70% surface RH), while increasing surface RH from 40% to 60% resulting in the transition of warm shallow clouds to deep convective
clouds due to the significant increase in CAPE values (Fan et al., 2007). The humidification of aerosols has also been suggested to be part of the source of reduced precipitation efficiency under conditions of high aerosol optical depth (AOD) especially in the tropics, while the reverse effect may occur in other regions (Boucher & Quass, 2013). The studies of Fan et al. (2009b, 2012b), involving multiple simulations of isolated DCC in different environments, suggests that vertical wind shear is a dominant factor in regulating aerosol effect on DCC, with strong wind shear enhancing evaporative cooling over condensational heating, leading to the suppression of convection. Likewise, it was found that weak wind shear allowed for enhanced condensational heating over evaporative cooling, leading to enhanced convection. The strong effect of CAPE on aerosol-DCC was reported by Lee et al. (2008), in which greater aerosol concentrations in strong CAPE environments resulted in stronger updrafts, increased condensate mass and increased precipitation, while a low CAPE environment produced less precipitation, due to the smaller cumulus clouds.

1.2 Deep Convective Clouds

Deep convective clouds (DCC) are important sources of precipitation and play a strong role in both regional and global circulation, with tropical convection being particularly significant (Arakawa, 2004). The strong updrafts within convective clouds can transport small cloud particles to the level of neutral buoyancy where they spread out to form the anvil associated with DCC (Folkins, 2002; Mullendore, 2005). The greater area coverage and lifetime persistence of the anvil cloud compared to the convective core makes the anvil cloud important to global energy balance and radiative transfer. This makes the study of deep convective clouds important for current and future climate research (Solomon et al., 2007; Rosenfeld et al., 2013). Convective intensity is the primary determiner of the depth, area, and lifetime of the resulting
anvil clouds (Futyan and Del Genio, 2007). However, observational and numerical studies of AIE suggest that changes to cloud microphysical processes can significantly modulate these macrophysical qualities (Fan et al. 2007a, 2010a, 2013; Min et al, 2009; Koren et al., 2010b; Li et al., 2011; Niu and Li, 2012; Saleeby et al., 2016). In addition, changes to precipitation vertical structure and convective/stratiform cloud ratios resulting from AIE can have a profound impact on the latent heat profiles of the atmosphere and subsequently regional and global circulations (Tao et al., 2006, 2007).

1.3 Ice Formation in DCC

Aerosols such as dust influence the character of individual clouds and storms, but evidence of a systematic effect on storm or precipitation intensity is still limited and ambiguous. Therefore detailed numerical models are required to understand the dynamical and microphysical changes and feedbacks that result in the observed effects of aerosols such as dust on DCC. However, the representation of DCC processes relevant to ACPI is still considered weak, due to the complex dynamics, thermodynamics and microphysics within DCC and the issue of some fundamental details of cloud microphysical processes still being poorly understood. This is particularly true with regards to ice and mixed-phase clouds (Boucher et al., 2013; Fan et al., 2016b). This low confidence is a result of the complex coupling between the processes controlling cloud and precipitation properties, which cover a wide range of spatial and temporal scales (Tao et al., 2012). As more than 50% of precipitation originates via ice-phase processes, it is critical to understand how aerosols impact ice formation and growth within deep convective clouds.
Earlier numerical studies of aerosol-cloud interactions tend to focus upon the action of soluble aerosols as CCN, with changes to ice formation resulting from the affected liquid processes only (Khain et al., 2005; Fan et al., 2009, 2012; Storer and van den Heever, 2013; Saleeby et al., 2016). However, DCC can also be sensitive to the aerosols that act as IN (Van den Heever, et al., 2006; Tao et al., 2007; Ekman et al., 2007). The study of Van den Heever et al. (2006) described the differing impacts of CCN and IN on convective clouds and subsequent anvil development. They found that increasing CCN concentration tended to reduce surface precipitation. Increasing IN concentration initially increased surface precipitation, but eventually reduced the total to less than the control case by the end of the simulation. Updraft intensity increased with the increased aerosol concentration due to stronger latent heat release resulting from the freezing of greater concentrations of supercooled water, but anvils were generally smaller and more organized. Ekman et al. (2007) studied the sensitivity of a continental storm to IN concentration and found that updrafts were enhanced due to added latent heat release from ice crystal depositional growth. The stronger updrafts enhanced homogeneous nucleation, increasing anvil cloud coverage and precipitation. Fan et al. (2010) compared the effects of CCN and IN on convection and precipitation and noted that the CCN effect is more evident in changing cloud anvil size, lifetime, and microphysical properties. IN was shown to have a small effect on convective strength, but the microphysical effects could still be significant. However it should be noted that Fan et al. (2010) did not have a prognostic IN treatment as what we have done for our studies.

1.4 Dust as Ice Forming Nuclei

The ice and liquid nucleating activity of any atmospheric aerosol particle is dependent on both its size and chemical composition. Silicate dust is one of the most prominent IN in the
atmosphere due to its significant concentration in the atmosphere (Prospero, 1999) and possession of favorable IN traits such as insolubility and crystallographic structure (Pruppacher and Klett, 1997). However it has also been reported that atmospheric processing, commonly termed ‘aging’ may leave a dust particle with a coating that impedes IN activity, by covering or destroying ice nucleating active sites (Tobo et al., 2012). It has also been reported that IN activity may only be significantly affected if the coating is chemically reactive with the underlying particle (Wex et al., 2014). Other studies have suggested that the IN activation of Saharan dust may actually be increased by atmospheric aging (Seifert et al., 2010; Boose et al., 2016). In the case of a soluble coating, dust particles may favor liquid nucleation over ice nucleation, their large size and soluble fraction allowing them to act as efficient CCN (Twohy et al., 2009). The presence of a soluble coating may also have an indirect effect on ice formation in liquid drop dependent freezing modes by depressing the freezing temperature of the drop due to added solute mass (Diehl and Wurzler, 2004). However, while soluble material is generally required for CCN activation, it has also been established that an insoluble particle with little to no soluble fraction may nucleate liquid drops under normal atmospheric conditions (Kumar et al., 2011; Karydis et al., 2013) allowing for dust to act as CCN with little effect on subsequent IN activity.
Chapter 2

Observational Studies of DCC-Dust Interactions

2.1 The Saharan Air Layer

Dust aerosols have been observed at significant concentrations even in remote locations far from their expected source regions (Prospero, 1999). They are predominately composed of insoluble silicate particles (Lohmann, 2002) which have been established to act as effective ice nuclei (IN, Pruppacher and Klett, 1997; Demott et al., 2003; Sassen et al., 2003; Boose et al., 2016) and/or cloud condensation nuclei (CCN; Twohy et al., 2009; Kumar et al., 2011; Karydis et al., 2013). The Saharan Air Layer (SAL; Prospero and Carlson, 1970; Carlson and Prospero, 1972) is an elevated layer of dry air between 850-500 hPa, often containing lofted dust particles. The SAL has been observed interacting with tropical cloud systems, such as tropical cyclones and mesoscale convective systems (MCS), and may impact their intensity and evolution (Karyampudi and Carlson, 1988; Dunion and Velden, 2004; Evan et al., 2008; Min et al., 2009; Zhang et al. 2009; Braun 2009; Lau et al., 2010; Carrios and Cotton, 2011; Cotton et al., 2012; Braun et al., 2013).

2.2 Overview of Case Study Observations

A trans-Atlantic dust outbreak of Saharan origin occurring 1-10 March 2004 (Morris et al., 2006) was subjected to a rigorous multi-sensor and multi-platform observational analysis (Min et al., 2009; Li et al., 2010; Min and Li, 2010; Li and Min, 2010; Min et al., 2014). The
interaction of this dust outbreak with a well-developed mesoscale convective system (MCS) resulted in strong effects on cloud microphysical processes. These observations included a comprehensive set of thermodynamic, dynamic and cloud microphysical observations of both the dust layer and the precipitation systems interacting with the layer. A special case on 8 March 2004 was selected for consideration, showcasing a well-developed MCS partially within the dust layer on the northern part of the system while the southern part remained in pristine maritime air (Min et al., 2009). Cloud ice particles were noted to be particularly abundant within the dust region, especially at temperatures conducive to heterogeneous ice nucleation. It was found that for a given convective strength, radar reflectivity values were lower in the dusty region compared to the dust-free region, with an increased prevalence of small hydrometeors in the stratiform regions. Larger convective reflectivity values within the dust-free region were accompanied by higher rain tops in both the convective and associated stratiform clouds, due to stronger convective activity. For the dust region, convective rain tops were lowered, while the stratiform rain top was higher than the dust-free region, so convective strength was not considered to be the dominant cause of the observed hydrometeor differences. Occurrence frequency plots of ice water path (IWP) with respect to cloud top temperature (CTT) showed a marked shift in the dust region away from temperatures where homogeneous drop freezing would be expected (-38°C), to warmer temperatures, suggesting that dust was acting as heterogeneous ice nuclei. A similar occurrence frequency plot of ice effective radius with respect to CTT illustrated a strong temperature dependence of particle size in the dust-free region that was contrasted by the nearly uniform particle size above -40°C and large change below -40°C in the dusty sector. Taken together, this was seen as evidence of the activity of dust acting as ice
nuclei, leading to changes in the cloud microphysical processes which shifted the precipitation size spectrum from heavy to light precipitation.

2.3 Longwave Indirect Effects of Dust on Ice clouds

The study by Min and Li (2010) assessed the impacts of the dust outbreak, described in Min et al. (2009), on ice cloud microphysical and macrophysical properties, focusing on associated thermal infrared radiative forcing of cloud systems over the Atlantic Ocean. A correlation was found between cloud effective temperature (CET) and aerosol optical depth (AOD) with a slope of +3.06 °C per unit AOD, suggesting that the minimum cloud effective temperature increases with AOD. This indicated that ice containing clouds were forming at warmer temperatures (lower altitudes) under higher AOD conditions in the study region. Additionally, there was no coherent correlation between AOD and IWP. As large IWP is assumed to be a result of thick ice clouds forming under strong convective activity with ample water vapor, IWP can be used as an indicator of dynamical and thermodynamical conditions. The study suggested that the observed tendency of CET with AOD cannot be attributed solely to changes in cloud dynamical and thermodynamical conditions. Differences between ice cloud properties and outgoing longwave radiation (OLR) under dusty and pristine conditions were compared statistically and related to the associated dynamical conditions using joint probability density functions (JPDF). CET was found to generally decrease with increasing IWP, but with a systematic shift evident in the JPDF for the dusty period compared to the pristine period. A greater number of ice clouds with IWP <300g/m² occurred in the dusty period formed at warmer temperatures (CET > -30 °C) compared to the pristine period. Ice clouds with IWP <300g/m² in the dusty period tended to have colder temperatures, implying that convective strength was increased for those clouds and suggesting that the lowered cloud top heights during the dusty
period were not a result of dynamical factors. Decreases in cloud top height can increase the OLR, which cools the atmosphere. OLR during the dusty period were substantially higher than the pristine period for both daytime and nighttime passes. Indirect effects from mineral dust resulted in an enhancement to OLR from 4 W/m to 16 W/m.

2.4 Impacts of Dust on Vertical Structure of Precipitation

The study of Li and Min (2010) examined the structure of vertical precipitation in the deep convective cloud system during the dust outbreak described in Min et al (2009) expanding upon that study with additional observations from TRMM and Aqua satellites, specifically using the passive microwave measurements from these instruments. Besides the separation of data into dusty and pristine periods, dynamical and thermodynamical conditions were constrained by separating precipitation into convective, stratiform, and warm rain types. The impacts of mineral dust were found to be dependent on rain type. For convective rain, the patterns of rainfall profiles above the dust layer are determined by updraft intensity, while within the dust layer, raindrop breakup was enhanced due to the presence of mineral dust and associated warm air (Dunion and Velden, 2004). For stratiform rain of similar storm height, precipitation was enhanced at altitudes above 6 km under dusty conditions. However, precipitation at lower altitudes was found to be reduced. Warm rain with similar storm height was found to be systematically weaker under dusty conditions. It was also found that the ratio of precipitation water to total amount of atmospheric hydrometeors, called the precipitation efficiency index, was smaller under dust laden conditions. It was indicated that a larger number of small-sized cloud particles formed in the dust-laden cloud system, which was explained as mineral dust acting as CCN and IN.
2.5 Impacts of Mineral Dust on Tropical Deep Convective Systems

The Study of Min et al. (2014) expanded upon the study of microphysical impacts of Saharan dust on deep convective ice clouds using multi-platform and multi-sensor observations. Comparison between dusty and non-dusty conditions supported the hypothesis that large concentrations of mineral dust increased the formation of more numerous ice particles through heterogeneous nucleation processes. In non-precipitating ice clouds, ice particle growth was limited due to competition between particles for available water vapor. In addition, less water vapor is transported to higher altitudes which reduces peak supersaturation values at those altitudes and consequently limits the formation of ice by homogeneous processes. This results in relatively small particle sizes and a narrower distribution of effective particle diameter, most notably at temperatures colder than -40°C. However, effective particle diameter of precipitating clouds in both dusty and non-dusty regions was found to be similar or smaller than particle sizes found within non-precipitating ice clouds. Particles can grow to larger sizes in precipitating clouds due to the stronger updraft strength and greater water vapor supply. It was theorized that these larger particles may have precipitated out of the cloud or settled out to lower parts of the cloud and were therefore undetectable by the optical sensors. Conversely, precipitating ice clouds were reported to have greater ice water paths under dusty conditions due to such clouds having sufficient water vapor supply to support significant particle growth. Mineral dust was suggested to invigorate convective activity as a result of increased latent heat release from the formation and subsequent growth of more numerous ice particles, which in turn enhances water vapor supply to deep convective precipitating clouds. This allows ice particles to be lifted to higher altitudes where they eventually evaporate, releasing water vapor and aerosols into the upper troposphere and lower stratosphere. Additional study, constraining the results based on
large scale sea surface temperature and associated dynamical conditions, illustrated that the observed changes to cloud effective diameter were a result of microphysical changes and not simply due to variability in large-scale dynamics and thermodynamics.
Chapter 3

Overview of the WRF-SBM Model and Experiment Design

3.1 Studying ACPI with Numerical Models

Large uncertainties also exist in ice nucleation parameterizations within numerical models (DeMott et al., 2010). However, comparison of model results with a well observed case study, such as the multi-platform and multi-sensor Min et al. (2009) study, can limit the impact of these uncertainties when analyzing results from numerical simulations. Radar reflectivity measurements provide a valuable insight into the microphysical impacts of aerosols, such as dust, on DCC when analyzed in conjunction with detailed numerical simulation results. However, radar reflectivity is sensitive to the number concentration, PSD, phase, density, fall rate, and spatial orientation of precipitation particles. These qualities are difficult to track accurately when a numerical model relies on the fixed PSDs frequently used within bulk microphysics schemes. The use of bin microphysics allows for explicit calculation of microphysical processes that affect cloud and precipitation formation and growth, such as diffusional growth processes, particle collisions, and particle sedimentation rates. In addition, the bin PSDs can be directly converted into radar reflectivity values for direct comparison with observations.

Ice formation in deep convective clouds may result from heterogeneous and/or homogeneous ice nucleation depending on the depth of the cloud and the chemical composition of the background aerosols. Heterogeneous ice nucleation can occur at temperatures between -
5°C and -38°C via the mechanisms of deposition, immersion, and contact freezing (Vali et al., 1985; Vali et al., 2015) when ice nuclei (IN) are present. Homogeneous ice nucleation involves droplet and aerosol haze particle freezing at temperatures lower than -38°C (Koop et al., 2000; Mohler et al., 2003; Ren and MacKenzie, 2005). Deep convection frequently shoots liquid drops up to the upper troposphere where the temperature is colder than -38°C, leading to strong homogenous droplet freezing. Therefore, a comprehensive handling both heterogeneous and homogeneous ice formation mechanisms must be incorporated into numerical simulations to gain a clearer understanding of ice formation in DCC.

Observations suggest that IN particles such as dust have a significant impact on the microphysical and macrophysical properties of DCC, but many numerical simulations rely on a relatively simple handling of IN particles and the associated heterogeneous ice formation mechanisms. Accurate simulations of ice formation processes in DCC require ice nucleation to be directly linked with IN concentration. In our studies, we add a prognostic IN variable to allow for the transport of IN particles by the wind field and the removal of IN by heterogeneous ice formation. We also update a set of heterogeneous and homogenous ice nucleation parameterizations within the WRF-SBM to connect ice nucleation with dust particles. Heterogeneous ice formation resulting from the updated immersion, contact, and deposition-condensation freezing schemes account for the full range of ice formation mechanisms active at temperatures between -5°C and -38°C. Detailed information on specific updates made to the WRF-SBM model has been provided below.

3.2 Weather Research and Forecasting (WRF) Model
Numerical simulations were undertaken using the WRF version 3.1.1 developed by the National Center for Atmospheric Research (NCAR) as described in Skamarock et al. (2008). WRF solves the fully compressible, non-hydrostatic Euler equations formulated on terrain following hydrostatic-pressure coordinates and the Arakawa C-grid. The model uses Runge-Kutta second- to sixth-order advection schemes in both horizontal and vertical directions. The fifth-order advection scheme is used in this study. The monotonic technique is employed for advection of scalar and moist variables. The cloud microphysical scheme employed within the study is described below.

3.3 Spectral Bin Microphysics (SBM)

The original SBM (Khain et al., 2004) solves a system of kinetic equations for the size distribution functions for 7 hydrometeor types: water droplets, ice crystals (plate, column, and dendrite), aggregates, graupel, and frozen drops/hail. An 8th size distribution function exists for CCN. Each size distribution is represented by 33 mass doubling bins, where the mass of a particle in each bin is twice the mass of a particle in the preceding bin. A fast version of the SBM (Fast-SBM) with four size distributions of water drops, low density ice (ice crystals and aggregates), high density ice (graupel and hail), and aerosol (CCN) was created in order to substantially reduce the computational costs (Khain et al. 2009, Fan et al 2012). Further details about the mechanics of the SBM can be found in Khain et al. (2004) and Fan et al. (2012) and will not be repeated here.

In order to examine IN impacts on clouds and precipitation, an additional prognostic variable for IN particle (dust in this case) number concentration was added to the model as detailed in Fan et al. (2014). We update the heterogeneous ice nucleation parameterizations in
the SBM (as detailed in the following section) to connect ice formation with dust particle concentrations. In this study, a dust layer located between 1km and 3km has been added to the dust case simulations, due to the presence of a similar dust layer in the observed case. The dust layer is initialized to cover the entire domain at model startup and thereafter is resupplied exclusively from the lateral boundaries by wind advection. The dust in the layer can serve as IN, CCN, or some fractional combination of the two by means of a simple partition set by the user depending on assumed or measured particle chemistry. This allows us to test the sensitivity of clouds within our model to a mixture of nuclei. In the current study, we have set the dust layer to supply IN exclusively in order to focus on the effects of heterogeneous ice formation and growth on subsequent cloud and precipitation processes. Therefore, these dust cases will represent the maximum potential IN effects on heterogeneous ice formation for a given dust number concentration. In addition to changes in nuclei concentration, temperature and moisture content within the dust layer can be modified to test the sensitivity of cloud and precipitation formation to such changes.

3.4 Ice Formation Parameterizations

The original SBM (Khain et al. 2004) included both homogeneous and heterogeneous ice formation, but did not directly connect ice formation to a prognostic IN variable. Liquid drop freezing for both homogeneous and immersion mechanisms was provided by Bigg (1953). Ice formation resulting from condensation and deposition freezing was provided by Meyers et al. (1992). Contact freezing was not included in the original SBM. In order to perform a study of aerosol impacts on heterogeneous ice formation, it is necessary to directly link ice nucleation rates to aerosol properties. The study of Fan et al. (2014) updated the available homogeneous freezing mechanisms and additionally implemented separate parameterizations into the SBM for
depositional, contact and immersion freezing, with ice formation in each of these schemes directly linked to the prognostic IN variable. In this study, we followed the Bigg (1953) for the homogeneous freezing of drops. The heterogeneous ice nucleation parameterizations employed within the simulations are detailed below.

3.4.1 Deposition and Condensation Nucleation

Currently there is no deposition and condensational nucleation parameterization connecting with aerosol properties and developed based on deep convective clouds. As noted in Meyers et al. (1992) it is difficult to distinguish the relative contributions of depositional and condensational freezing in a parcel, since both form similarly sized small ice crystals, despite the different mechanisms of vapor to ice in the former and condensation followed immediately by freezing in the latter case. However, studies suggest that small ice crystals formed in the -5°C to -10°C temperature range can have a large impact on subsequent ice formation at higher altitudes (Ackerman et al. 2015; Hiron and Flossman, 2015; Lawson et al., 2015). A depositional-condensational scheme would allow for these small ice crystals to form in this specific temperature range. To link depositional and condensational freezing with aerosols, we follow the implementation of van den Heever et al. (2006), updated from the Meyers et al. (1992) parameterization. The number of ice crystals generated by depositional-condensational nucleation ($N_{\text{dep}}$) is proportional to the IN number concentration ($N_{\text{IN}}; \text{l}^{-1}$) within the grid cell by Eq. (1):

$$N_{\text{dep}} = N_{\text{IN}} F_M,$$

(1)
where $F_M$ (unitless) is the function of the depositional-condensational nucleation by Meyers et al. (1992) that represents the fraction of the maximum available IN ($N_{id}; \text{l}^{-1}$) concentration that may be activated for the given conditions as calculated in Eq. (2):

$$N_{id} = \exp\{-6.39 + 0.1296[100(S_i - 1)]\},$$

(2)

with $S_i$ being the supersaturation over ice. The value of $F_M$ is equal to 1 for conditions at ice supersaturation of 40%, at which point all IN are activated, and is equal to 0 when supersaturation over ice is negative. The initial size of an ice crystal formed by this scheme is assumed to be 2.5 μm in radius and is assigned to the smallest ice size bin.

### 3.4.2 Immersion Freezing Nucleation

As stated above, the immersion freezing mechanism in the original SBM uses the parameterization of Bigg (1953), which is temperature-dependent only. To provide an aerosol-based immersion freezing scheme, we have incorporated the parameterization of DeMott et al. (2015), which was implemented by Fan et al. (2014) (cited as DeMott et al. (2013) in Fan et al. (2014) due to DeMott et al. (2015) not yet being published). The DeMott et al. (2015) immersion freezing activation number is parameterized as in Eq. (3):

$$N_{imm} = (CF)(N_{IN})^{(\alpha(273.16-T_k)+\beta)}\exp(\gamma(273.16-T_k)+\delta)$$

(3)

$CF$ is an instrumental correction factor with a value of 3. Coefficients $\alpha$, $\beta$, $\gamma$, and $\delta$ are 5.95E-5, 1.25, 0.46, and -11.6, respectively, representing mineral dust particles (Demott et al., 2015) $T_k$ is the cloud temperature in degrees Kelvin, $N_{IN}$ is the number concentration of total aerosol particles with diameter larger than 0.5 μm, and $N_{imm}$ is the maximum number of immersion ice possible in the given temperature range. Liquid drops are consumed over the size spectrum
starting with the largest sizes down to the smallest until the minimum of $N_{imm}$ or drop number is reached. According to Yin et al. (2005), drops with a radius smaller than 79.37 μm will be frozen to pristine ice crystals, otherwise graupel is formed.

### 3.4.3 Contact Nucleation

We have also adopted the contact freezing parameterization of Muhlbauer and Lohmann (2009), which is based on Cotton (1986) and Young (1977). In this parameterization, contact freezing is a result of the collision of supercooled liquid water drops and IN particles due to Brownian motion. The contact freezing rate is therefore proportional to the drops' radius and number concentration. It is also proportional to the model’s IN number concentration and Brownian diffusivity in air. Unlike Muhlbauer and Lohmann (2009) who calculated the freezing rate for the sum of all drops, we perform the calculation in this study for each spectral bin of drops. Then, the contact freezing activation number ($N_{crt}$; $l^{-1}s^{-1}$) for each individual size bin is represented by Eq. (4):

$$N_{crt} = 4\pi r_c N_C D_k N_{IN},$$

(4)

where $r_c$ (m) and $N_C$ (m^{-3}) is radius and number concentration of drops in the bin, respectively. $D_k$ is the dust aerosol Brownian diffusivity (m^{2}s^{-1}), and is parameterized by Eq. (5):

$$D_k = \frac{k_BT C}{6\pi \eta r},$$

(5)

$D_k$ is a function of the Boltzmann constant $K_B=1.28 \times 10^{-23}$ m^{2} kg s^{-2} K^{-1}, $T$ is the air temperature, $r$ is the dry dust aerosol median radius, $\eta$ is the viscosity of air and $C$ is the Cunningham slip correction factor. The viscosity of air depends on temperature, as calculated by Eq. (6):
\[ \eta = 10^{-5}[1.718 + 4.9 \times 10^{-3}(T - 273.15) - 1.2 \times 10^{-23}(T - 273.15)^2]. \quad (6) \]

The Cunningham slip correction factor is calculated by Eq. (7):

\[ C = 1 + 1.26 \lambda \frac{T}{r} \frac{10^{1.25}}{p} \frac{T}{273.15}, \quad (7) \]

with the molecular mean free path length of air \( \lambda = 0.066 \mu m \), \( r \) is the dry aerosol radii, and \( p \) is the pressure. To simplify the calculation, the contact freezing number is the available dust number concentration \( N_{IN} \), with freezing efficiency of 1. Upon freezing, drops with a radius smaller than 79.37 \( \mu m \) will be frozen to pristine ice crystals, larger drops will be frozen as graupel.

It should be noted that currently there is no ice nucleation parameterization specifically developed for DCC, and the understanding of ice nucleation for DCC is still very limited. The best we can do for model simulations at this time is to employ the currently-available ice nucleation parameterizations for connecting with dust particles, evaluate our baseline simulation with observations, and carry out model sensitivity tests based on the validated case simulation to understand the dust impacts and associated mechanisms.

### 3.5 Radar Reflectivity Calculations

The liquid and frozen hydrometeor PSDs calculated by SBM can be easily converted into radar reflectivity values, providing a bridge for the comparison of model simulated microphysical parameters with observable variables. For our study, we calculate radar reflectivity directly from the model’s PSD for each of the individual hydrometeor species using the spherical particle approximations of the Rayleigh scattering equations suggested by Ryzhkov et al. (2011). Reflectivity is calculated for each bin and then summed over the entire PSD to
obtain the total reflectivity. The general equation for snow and graupel reflectivity is represented by Eq. 8:

\[ Z = \left( \frac{\rho_{s,g}}{\rho_i} \right)^2 \frac{|K_i|^2}{|K_w|^2} \int_0^\infty D^6 N(D) dD \]  

(8)

Where \( N(D) \) is the number concentration per cubic meter of snow(graupel) particles of Diameter (D) in millimetres. Density of snow or graupel is represented by \( \rho_{s,g} \), while \( \rho_i \) is the density of solid ice. \(|K_i|^2 \) and \(|K_l|^2 \) represent the dielectric factors of solid ice and liquid water, respectively. When calculating the reflectivity for liquid drops, the two leading ratios are equal to 1, but otherwise the equation is the same. The density relationship in the leading ratios can be expanded and simplified into a constant times the snow(graupel)-liquid density ratio, following Smith (1984) and Fovell and Ogura (1988) as in Eq. 9:

\[ \left( \frac{\rho_{s,g}}{\rho_l} \right)^2 \frac{|K_l|^2}{|K_w|^2} = \left( \frac{\rho_{s,g}}{\rho_l} \right)^2 \left( \frac{\rho_l}{\rho_i} \right)^2 \frac{|K_l|^2}{|K_w|^2} = 0.224 \left( \frac{\rho_{s,g}}{\rho_l} \right)^2 \]  

(9)

Where \( \rho_l \) represents the density of liquid. This is then substituted into Eq. 8 to yield Eq. 10:

\[ Z = 0.224 \left( \frac{\rho_{s,g}}{\rho_l} \right)^2 \int_0^\infty D^6 N(D) dD \]  

(10)

The reflectivity values calculated for liquid drops, snow and graupel are then added together to obtain the total reflectivity, which is converted to dBZ by Eq. 11:

\[ Z_{dBZ} = 10 \log(Z_{total}) \]  

(11)
3.6 Experiment Design

3.6.1 Initial and Boundary Conditions

In our study, we have simulated the 08 March 2004 MCS described in Min et al. (2009), using realistic initial and boundary conditions, which are provided by the 1° x 1° 6-hourly National Centers for Environmental Prediction (NCEP) global final analysis dataset. Each simulation was run for 33 model hours, beginning at 18Z 07 March 2004. Due to the SBM being too computationally expensive to run at coarse resolutions, the SBM provides microphysics for only the smallest domain while bulk microphysics are used within domains 1-3. The specific bulk microphysics scheme and other model parameterizations selected for use within our study are provided in Table 1. The initial vertical profile of domain averaged relative humidity indicates moist (>60% RH) air below 6km and drier air (<50% RH) above 6km, while the profiles of horizontal winds evidence weak (<5 m/s) to relatively weak (<10m/s) wind shear below 7km, following the criteria used by Fan et al. (2009b). After the model’s 6 hour spin up time, a dry air layer corresponding to the SAL enters the domain via the NCEP-FNL boundary conditions and is present for the duration of the simulation.
Table 3-1: WRF model parameterizations selected for use in case study simulations.

<table>
<thead>
<tr>
<th>Parameterization</th>
<th>Selected option</th>
</tr>
</thead>
<tbody>
<tr>
<td>Microphysics</td>
<td>Domain 1,2,3: Thompson (Thompson et al., 2008); Domain 4: SBM (cited in text)</td>
</tr>
<tr>
<td>Cumulus</td>
<td>Domain 1,2: Kain-Fritsch (Kain, 2004)</td>
</tr>
<tr>
<td>LW Radiation</td>
<td>Rapid Radiative Transfer Model (Mlawer et al., 1997)</td>
</tr>
<tr>
<td>SW Radiation</td>
<td>Dudhia scheme (Dudhia, 1989)</td>
</tr>
<tr>
<td>PBL</td>
<td>MYNN2(Nakanishi and Niino, 2006)</td>
</tr>
<tr>
<td>Surface layer</td>
<td>MM5 similarity (Zhang and Anthes, 1982)</td>
</tr>
<tr>
<td>Land surface</td>
<td>RUC LSM (Smirnova et al., 1997)</td>
</tr>
</tbody>
</table>

3.6.2 CCN and IN distributions

To simulate a pristine maritime environment, the CCN number is set to a uniform value of 300 cm$^{-3}$ below 2km with the CCN number being reduced exponentially from this value as height increases above 2km. The initial IN distribution is set to be vertically uniform at 0.01 cm$^{-3}$ for the Clean case. The dust cases add a progressively larger number of IN to the Clean case’s background value in a dust layer located vertically between 1km and 3km. The initial vertical profiles of CCN and IN are shown in Figure 3-1. As dust particles are known to be potential CCN (Twohy et al., 2009; Kumar et al., 2011; Karydis et al., 2013) as well as IN, the dust layer includes a partition allowing a fraction of the total dust number to activate as CCN. In the current study, we have focused on the sensitivity of cloud and precipitation formation processes to additional IN activation exclusively. Therefore, the CCN/IN partition is set to exclusively
provide IN in order to provide an upper limit on the effects of IN activation for a given IN number concentration. To prevent the CCN and IN fields from being diluted due to the inflow of air from the lateral boundaries, the CCN and IN numbers of the outer five grid cells (i.e., the boundary points) on each side of domain 4 are set to the initial values throughout the integration period.

The initial dust particle number concentrations within the dust layer are: 0.12 cm$^{-3}$ (case D.12), 1.2 cm$^{-3}$ (case D1.2), and 12 cm$^{-3}$ (case D12), respectively. These values were selected based on aerosol measurements (Table 2) taken during the trans-Atlantic Aerosol and Ocean Science Expeditions (AEROSE) experiment (Morris et al., 2006) for dates coinciding with the observational study of the March 2004 dust outbreak detailed in Min et al. (2009). The dust loading was assumed to be the difference in the aerosol number of the dusty and pristine periods. Only aerosol particles with a radius greater than 0.5 microns were considered when taking this difference, due to the smaller aerosol sizes being more prevalent during the pristine period compared to the dusty period. This size range is consistent with the study of DeMott et al. (2015) for ice nucleating particles. The resulting dust number was multiplied by an activation fraction suggested by Niemand et al. (2012) for Saharan dust to arrive at the number used for case D.12. Other studies have suggested that dust related IN numbers greater than 1.0 cm$^{-3}$ are possible (DeMott et al., 2003; Sassen et al., 2003; Ansmann et al., 2008), so two additional dust cases with IN numbers one (D1.2) and two (D12) orders of magnitude greater than the initial D.12 case were included in the study.
Figure 3-1: Vertical profiles of domain initial and boundary grid point number concentrations of CCN (dashed) and IN (solid) number concentrations, per cubic centimeter (pcc, Log\(_{10}\)).

Colors represent Clean (black), D.12 (blue), D1.2 (red), and D12 (green) cases, respectively.

Table 3-2: Ship observed aerosol number concentrations from the AEROSE campaign corresponding to the March 2004 Saharan dust outbreak.

<table>
<thead>
<tr>
<th>March 2004 Ship Observed Aerosol Number</th>
<th>Radius 0.3-0.5</th>
<th>0.5-1</th>
<th>1-5</th>
<th>5-10</th>
<th>10-25</th>
<th>25</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dust-free</td>
<td>108.6</td>
<td>10.5</td>
<td>2.36</td>
<td>0.1029</td>
<td>3.00E-4</td>
<td>5.50E-8</td>
</tr>
<tr>
<td>Dust</td>
<td>87.32</td>
<td>34.7</td>
<td>7.557</td>
<td>0.3537</td>
<td>1.45E-3</td>
<td>7.41E-6</td>
</tr>
</tbody>
</table>

cm\(^{-3}\), micron
3.6.3 Cloud Type Classification Criteria

Additional criteria used to select subsets of the data for the purpose of our analysis are as follows. Cloudiness within an individual 3D grid cell was determined by the sum of all condensates within it exceeding a $10^{-6}$ kg/kg threshold value, following the definition used in Fan et al. (2013). Cloud top was determined, from the top level of the model down to the surface, as the highest level with at least two consecutive levels exceeding the cloudiness threshold, which was intended to limit the influence of very thin clouds on the resulting analysis. To sort results by precipitation regime, we adapt the definitions of Fan et al. (2013) for convective and stratiform precipitation, with each vertical column classified as a single precipitation type only. Separation of clouds into convective and stratiform precipitation regimes when analysing results was intended to minimize the variability related to different cloud evolution stages.

For all precipitating clouds, surface rain rates must exceed 0.05 mm/hr. Convective precipitation is classified as precipitating column with vertical motion exceeding a 1m/s threshold and cloud thickness of 8km or greater. In addition, the grid points immediately surrounding the column matching these criteria are also classified as convective to capture the transition between convective and stratiform clouds. The non-convective precipitating columns are classified as stratiform by the presence of ice-phase precipitation in the column. Non-precipitating columns with a cloud layer thicker than 1km and both cloud top and cloud bottom temperatures colder than 0°C are classified as anvil clouds. Precipitating columns with cloud top temperatures warmer than freezing are classified as rain producing warm clouds.
Chapter 4

Ice Formation and Cloud Properties

Abstract

To better understand the impacts of dust aerosols on deep convective cloud (DCC) systems reported by previous observational studies, a case study in the tropical eastern Atlantic was investigated using the Weather Research and Forecasting (WRF) model coupled with a Spectral Bin Microphysics (SBM) model as a two part study. In the first part, a detailed set of ice nucleation parameterizations linking ice formation with aerosol particles have been implemented in the SBM. Increasing IN concentration in the dust cases results in the formation of more numerous small ice particles in the heterogeneous nucleation regime (between -5°C and -38°C) compared to the background (Clean) case. Convective updrafts are invigorated by increased latent heat release due to depositional growth and riming of these more numerous particles, which results in increased overshooting and higher convective core top heights. Homogeneous ice formation is reduced in the dust cases as IN concentration is increased, due to more liquid drops converting to ice by freezing or riming before reaching -38°C and reduced peak supersaturation values from increased diffusional growth. Local IN activation in the stratiform regime contributes to increased cloudiness in the heterogeneous nucleation regime. A greater number of large snow particles form in the dust cases, which are transported from the core into the stratiform regime and sediment out quickly. Together with reduced homogeneous ice formation, fewer particles form within and/or are transported into the anvil regime. This shifts
the stratiform/anvil cloud occurrence frequency to warmer temperatures and reduces anvil cloud extents.

4.1 Introduction

Deep convective clouds (DCC) are important sources of precipitation and play a strong role in both regional and global circulation, with tropical convection being particularly significant (Arakawa, 2004). Observational and modeling studies of DCC have shown different results relating to the effect of aerosol on convection and precipitation, indicating that aerosol may either enhance or suppress convection and precipitation depending on aerosol concentration and environmental conditions (Khain and Pokrovsky, 2004; Khain et al., 2004, 2005, 2008; van den Heever et al., 2006; Fan et al., 2007b; Lee et al., 2008; Min et al., 2009; Min and Li, 2010; Li and Min, 2010; Min et al., 2014; Altaratz et al., 2014). A trans-Atlantic dust outbreak of Saharan origin occurring 1-10 March 2004 (Morris et al., 2006) was subjected to a rigorous multi-sensor and multi-platform observational analysis (Min et al., 2009; Li et al., 2010; Min and Li, 2010; Li and Min, 2010; Min et al., 2014). The interaction of this dust outbreak with a well-developed MCS resulted in strong effects on cloud microphysical processes. Small ice particles were abundant in the heterogeneous nucleation regime in the dusty region. The size spectrum of the vertical precipitation structures was shifted from heavy to light precipitation (Min et al., 2009; Li and Min, 2010). Substantial changes to cloud top distributions and precipitation profiles resulted from a change in the partition between homogeneous and heterogeneous ice formation processes under dusty conditions. Such macrophysical changes in the cloud systems resulted in substantial thermal infrared radiation cooling of up to 16 Wm$^{-2}$ (Min and Li, 2010). The reported changes to cloud top distribution and the partition between homogeneous and heterogeneous ice formation differ from those described by studies focusing on the CCN activation of aerosols.
Quantifying the effects of dust and other aerosols on deep convective clouds (DCC) is difficult due to the complex coupling between microphysical, dynamical and radiative processes involved, especially regarding ice and mixed phase processes (Tao et al., 2012; Boucher et al., 2013). The action of dust as IN is particularly important since precipitation formation is efficient through ice-related processes. Therefore the use of a detailed numerical model is crucial in untangling the effects of microphysics from dynamical and thermodynamical processes.

4.2 Results

Min et al. (2009) reported a unique case of a mature MCS partially under the effects of a Saharan dust outbreak. They noted distinct changes to cloud microphysical and macrophysical properties when comparing the dusty and dust-free sectors of the MCS. Large-scale meteorological conditions drive the initial cloud formation and growth processes which are then modulated by aerosol indirect effects on cloud microphysical processes. Figure 4-1 describes the locations of the four model domains, displaying the Atmospheric Infrared Sounder (AIRS) retrieval (Figure 4-1a) and the domain 1 model simulated precipitable water averaged over the duration of the simulation (Figure 4-1b). The large scale patterns of precipitable water are well reproduced by the model, although we note that the magnitude is slightly overestimated over the African continent and underestimated over the southern Atlantic compared to observations. Despite this, the magnitude in the location of our smallest domain is well reproduced, suggesting that the meteorological conditions in our region of interest are represented sufficiently well.
**Figure 4-1:** (a) AIRS total precipitable water averaged 07-09 March 2004, boxes denoting location of the three nested domains. (b) Domain 1 model output precipitable water averaged 07-09 March 2004.

### 4.2.1 Overall Microphysical and Macrophysical Changes

Our numerical simulation of the case study MCS tests the effects of different number concentrations of IN on heterogeneous and homogeneous ice formation and the subsequent changes to cloud microphysical and macrophysical properties by comparing a Clean (background) case with three dust cases (D.12; D1.2; D12) as described in chapter 3. Min et al. (2009) and Min and Li (2010) observed increased ice water path in dust affected clouds with cloud top heights colder than -15°C but warmer than -38°C, suggesting a significant source of IN for heterogeneous ice formation. Figure 4-2a describes the vertical distribution of total (liquid+ice) water content (TWC). The TWC is averaged horizontally over all cloudy data points.
and is temporally averaged from model hour 6 to hour 33. In our simulations, increasing IN concentration progressively increases TWC at temperatures below 0°C (~ 5km) due to increased heterogeneous ice formation. When heterogeneous ice formation is sufficiently large, such as in the D12 case, particle growth rates are reduced due to competition between individual particles for available water vapor. This reduced growth rate is visible in the initially positive slope of TWC at temperatures below 0°C in the D12 case compared to the other cases as well as the reduced rain rates near 6km (Figure 4-2c). Diffusional growth of these more numerous particles results in greater latent heat release in the dust cases. This in turn increases the relative fraction of strong (>1m/s; Figure 2d) updrafts occurring at temperatures where heterogeneous ice formation is significant.

Li and Min (2010) reported that stratiform precipitation at temperatures below 0°C is enhanced for a given storm height in the dusty region. However, precipitation at warmer temperatures was reduced. The formation of large precipitation particles is generally a result of particle collisions rather than diffusional growth and is affected by changes in the particle size distribution (PSD). The formation of smaller but more numerous ice particles in the heterogeneous nucleation regime reduces the efficiency of particle collision and collection processes in the dust cases. This shifts the formation of precipitation to colder temperatures within the heterogeneous nucleation regime as particles grow in size during vertical transport. The average fraction of precipitation water content to TWC (Figure 4-2b) is increased in the dust cases compared to the Clean case at temperatures below 0°C. This is consistent with the findings of Li and Min (2010). While the precipitation mass fraction is not significantly reduced at temperatures above 0°C compared to the Clean case, the TWC is reduced overall. The profiles of average vertical rain rates (Figure 4-2c) indicate reduced near surface rain rates and increased
rain rates in the heterogeneous nucleation regime which is consistent with the observations of Min et al. (2009) and Li and Min (2010).

Figure 4-2: Domain averaged vertical profiles of in-cloud (a) total water content (TWC), (b) fraction of precipitation water content to TWC, and (c) ice+liquid rain rate. (d) Relative fraction of strong updraft (> 1m/s) grid points to total updraft grid points at each altitude. Colors represent Clean (black), D.12 (blue), D1.2 (red), and D12 (green) cases, respectively.
Aerosol indirect effects on cloud microphysical processes can result in a cloud top distribution that is higher or lower than would be expected for a given convective dynamical intensity. Figure 4-3 describes the changes to cloud top distribution in each of the three dust cases with respect to the Clean case. The cloud top distribution in Figure 4-3 combines all cloud types together to describe the overall macrophysical changes due to increasing IN concentration. We determine the cloud top by selecting the highest vertical model level in each column that exceeds the 1.e-6 kg/kg cloudiness threshold value. While this does not take multiple cloud layers into account, it is similar to the top-down view of clouds observed by many satellites. Figure 4-3a describes the time series of cloud top occurrence frequency for the Clean case. The percentage at each model time represents the horizontal sum of all cloud tops occurring at a given model level divided by the total horizontal and vertical sum of cloud tops occurring at that specific model output time. Figure 4-3b through Figure 4-3d describe the dust case minus Clean case difference of cloud top percentage. Increasing IN concentration in our simulations results in the overall cloud top height distribution shifting to lower altitudes (warmer temperatures). This is consistent with the findings of Min and Li (2010) in which higher AOD values were correlated with warmer cloud effective temperature. These macrophysical changes in cloud top distribution were noted to result in a strong cooling effect of thermal infrared radiation.
Figure 4-3: Time series of percentage of cloud tops occurring at each altitude for the (a) Clean case and the associated dust case minus Clean case differences plots for the (b) D.12 - Clean, (c) D1.2 - Clean, and (d) D12 - Clean cases.

The cloud system transitions from shallow to deep convection between model hours 6 to 12. The majority of cloud tops occurring before hour 10 are warmer than -5°C. Therefore, the temperature and supersaturation conditions within these clouds are not sufficient for IN to activate and form ice crystals. Hence, the effects of increasing IN are limited during this time period. After the transition to deep convection, the cloud top distribution is shifted to lower
altitudes (warmer temperatures) between model hours 12 to 24. Cloud tops occur less frequently above 15km and more frequently between 12km and 13km as a result of the changes in the partition between homogeneous and heterogeneous ice formation. This is most pronounced in the D1.2 and D12 cases which both feature significant increases in heterogeneous ice formation compared to the Clean case. The numerous ice crystals that form when large concentrations of IN are activated compete for available water vapor during diffusional growth. The consumption of the cloud’s available water vapor reduces peak supersaturation at colder temperatures and suppresses homogeneous ice nucleation. We note that the shift in cloud top distribution is not linear with increasing IN number concentration. While both the D1.2 and D12 cases feature lowered clouds (hour 12-24), the differences in the D12 case are not as pronounced as in the D1.2 case. This is a result of stronger updrafts (Figure 4-8) and greater concentrations of small cloud ice particles (Figure 4-7) in the D12 case compared to the D1.2 case. The small ice particles remain near the cloud top after the larger particles sediment out, yielding a higher cloud top distribution relative to the D1.2 case. After model hour 24, the cloud top distribution is significantly lowered in the D12 case compared to the D1.2 case. The greater condensate mass of the D12 case allow more large snow particles to form (Figure 4-7d), which sediment out more quickly compared to the D1.2 case (Figure 7c). The small IN number in the D.12 case results in a cloud top distribution that is different from both the D1.2 and D12 cases. From model hours 20 onwards, The D.12-Clean case difference plot suggests that higher cloud tops are occurring compared to the other cases. However, average convective updrafts are slightly weaker during this time period compared to the Clean case (Figure 4-8a). This suggests that cloud microphysical changes are the cause of the higher clouds in the D.12 case. Specifically that
particle sizes are smaller, allowing for increased vertical transport and slower sedimentation rates.

### 4.2.2 Radar Reflectivity

Min et al. (2009) used contoured frequency by altitude (CFAD) plots to describe the observed changes to convective and stratiform radar reflectivity between the dusty and dust-free regions. They noted that radar reflectivity at temperatures above 0°C was reduced in the dusty region in both the convective and stratiform regimes. At temperatures below 0°C, convective reflectivity was reduced in the dusty regions while stratiform reflectivity was increased. Min et al. (2009) performed a sensitivity test to differentiate the effects of dynamics on hydrometeor growth and precipitation formation from the microphysical effects of dust. The sensitivity test revealed that, in the absence of dust, relatively stronger convective intensity also resulted in higher stratiform reflectivity values. This indicated that the reduced reflectivity in the convective regime and increased stratiform reflectivity observed in the dust sector were a result of changes to microphysical processes rather than dynamics. These microphysical changes were suggested to be a result of increased heterogeneous ice formation, which delayed the formation of large precipitation particles in the convective regime until sufficient growth occurred during transport into the stratiform regime to pass the minimum reflectivity threshold (Min et al., 2009).
**Figure 4-4:** Contoured frequency by altitude diagrams (CFAD) of model simulated convective (top row) and stratiform (bottom row) reflectivity. Columns: Clean, D.12-Clean, D1.2-Clean, D12-Clean cases, respectively.

The use of bin microphysics in our WRF-SBM model allows us to simulate radar reflectivity directly from each respective hydrometeor’s size distribution. These PSDs evolve naturally within the model during cloud formation and growth. Therefore, a more accurate depiction of microphysical changes to precipitation formation and the associated changes to radar reflectivity is possible in comparison with the fixed hydrometeor PSDs used in bulk radar simulators. To compare the observations of Min et al. (2009) with our results, we have recreated similar CFAD plots using model derived reflectivity. Figure 4-4 describes the radar reflectivity CFADs of the convective and stratiform regimes for the Clean and three dust cases. As IN concentration is increased in the simulations, convective reflectivity at temperatures above 0°C is
decreased. Likewise, stratiform reflectivity at temperatures below 0°C is increased in the dust cases. These changes suggest that increased heterogeneous ice formation is significantly affecting the formation of precipitation sized particles consistent with the hypothesis of Min et al. (2009). Min et al. (2009) report that reflectivity is reduced over these temperature ranges in the respective cloud regimes. We will present a more in-depth discussion of the effects of dust on precipitation formation processes in Chapter 5.

4.2.3 Effects on Ice Nucleation

The convective core is the primary determiner of cloud macrophysical properties such as cloud top height and anvil cloud area (Futyan and Del Genio, 2007). However, changes to cloud microphysical processes resulting from AIE will modulate these macrophysical properties differently depending on the aerosol type. In our numerical simulations, increasing IN in the dust cases increases the total number of new ice crystals forming in the heterogeneous nucleation regime between -5°C and -38°C (Figure 6b to Figure 6d). This affects the vertical distribution of ice particles by changing the locations of initial ice formation and subsequent growth. Figure 4-5a and Figure 4-5b describe the vertical distribution of ice particles formed by the model’s heterogeneous and homogeneous ice formation schemes in the convective and stratiform cloud regimes, respectively. The ice formation number at each vertical level is summed horizontally and with respect to time for each cloudy pixel in the appropriate cloud regime and is represented by a log$_{10}$ value. Figure 4-5c describes the vertical distribution of residual (non-activated) IN number concentration in the convective cloud regime. This value is averaged over all convective cloud data points and temporally over the duration of the simulation. Increasing IN concentration in the convective core results in significant increases in ice formation between -5°C and -15°C. Ice formation in this temperature range can deplete available IN (Figure 4-5c) and reduce
heterogeneous ice formation between -15°C and -38°C. This depletion effect is substantial between ~7km and 11km in the D.12 and D1.2 cases. When the IN concentration is sufficiently high, such as in the D12 case, depletion is not as significant as in the other cases and ice formation is significantly increased over the majority of the -5°C to -38°C temperature range. At the -38°C threshold, ice formation number is progressively reduced as IN number is increased, which suggest that clouds are glaciating at warmer temperatures. The reduction of homogeneous freezing is due both to fewer liquid drops crossing the -38°C threshold (Figure 4-6j to Figure 4-6l) and reduced peak supersaturations resulting from increased ice growth at temperatures above -38°C. Finally, we note that stratiform ice formation is also increased in the dust cases compared to the Clean case. The increase, while not as large as in the convective core, contributes to increased cloudiness in the stratiform regime between -5°C and -38°C by increasing local concentrations of small, slow-falling ice crystals.
4.2.4 Convective Regime Hydrometeor Number Concentration and PSDs

Changes to the location and number concentration of initial ice particle formation affect the vertical distribution of ice and liquid hydrometeors in several ways. Figure 4-6a to Figure 4-6d describes the time evolution of convective averaged ice and snow particle number concentration. Increasing IN concentration results in a greater number of ice/snow particles in the heterogeneous nucleation regime and a corresponding reduction within the cloud column at temperatures below -38°C. This indicates that the reduced homogeneous ice formation number noted in Figure 4-5a is not counteracted by the transport of a similar number of particles from temperatures warmer than -38°C. While more particles are formed in the heterogeneous
nucleation regime between -5°C and -38°C compared to the Clean case, there are also more opportunities for these particles to collide and be incorporated into larger particles. For example, more frequent riming of ice and snow particles in the dust cases increases the formation of graupel (Figure 4-6e – Figure 4-6h). More frequent riming in turn reduces the average number of liquid drops in the convective regime at temperatures colder than -5°C (Figure 4-6i – Figure 4-6l). While dust only activates as IN and not CCN in our simulations, average liquid drop number at temperatures above -5°C is affected by the more numerous ice particles forming in the heterogeneous nucleation regime and subsequently melting after falling into warmer temperatures. Small ice particles melt into small drops that may evaporate, while large drops formed from melted snow/graupel collect smaller drops by collision-coalescence or may break up into smaller drops themselves.
Figure 4-6: Convective cloud averaged profiles (Height vs time) and dust case minus Clean case difference plots of ice number concentration (top row), graupel number concentration (middle row), and liquid number concentration (Bottom row). Columns: Clean (a,e,i), D.12 - Clean (b,f,j), D1.2 - Clean (c,g,k), and D12 - Clean (d,h,l) cases.

Increasing the total number of ice particles formed in the heterogeneous nucleation regime affects the PSD in two ways. First, available water vapor is partitioned over a greater number of smaller particles. Second, these smaller particles are less efficient at colliding with other particles. Both effects reduce the growth rates of the individual particles and shift the PSD to smaller sizes overall. The SBM allows us to examine the effects of dust on the PSD of the different hydrometeors without creating an arbitrary distinction between cloud and precipitation sizes particles. Dust related changes to the bin PSD of each hydrometeor type are described in
Figure 4-7. The provided radii values of the represented hydrometeor species are taken from the pre-calculated bin radii values used by the model, which are based on assumed particle densities and the mass doubling relationship between the individual bins. Contour values represent \( \log_{10} \) values of bin number concentration. The difference plots likewise describe the relative change of these \( \log_{10} \) values, representing \( \log_{10}(\text{Dust/Clean}) \) values. As dust in our study acts as IN exclusively and not CCN, we focus our discussion on the -5°C to -38°C degree range conducive to heterogeneous nucleation and freezing. Since dust in nature can also act as effective CCN and may therefore be removed from the system by warm rain processes before freezing occurs, these results should be interpreted as an upper range of IN effects for a given dust number concentration.

Figure 4-7a, Figure 4-7e, and Figure 4-7i describe time series of the PSD averaged over convective data within the -5°C to -38°C temperature range for ice/snow, cloud/rain drops, and graupel. The rest of Figure 4-7 shows the differences between the three dust cases and the Clean case. The addition of dust to the DCC system under strong convection produces an initial burst of ice formation covering the range of the PSD. In the D.12 and D1.2 cases, this is followed by a reduction in the small crystals and an increase in larger crystals and snow between hours 12 and 24. IN concentration has been depleted during this time period, which reduces the formation of small ice crystals. Existing ice crystals grow by particle collection into snow, hence the upwards slope in the difference contours between hour 12 and 24. When IN concentration is sufficiently large (D12 case) depletion is not as significant and small ice crystals form over the duration of the simulation. The liquid PSD describes an enhancement to the largest drop sizes that could be the result of increased collision-coalescence and/or the vertical transport of recently melted large ice particles. The middle size range of the liquid PSD is reduced though the duration of the
simulation, corresponding with the enhanced population in the graupel PSD. The formation of
graupel in our model occurs by two distinct mechanisms, direct freezing of large liquid drops, by
the homogeneous or immersion/contact freezing mechanisms, and collisions between liquid and
ice particles. There is evidence of increased large drop freezing, as seen by the enhancement to
the largest bin sizes in the graupel PSD. However, the majority of graupel particles are being
formed by riming, as seen by the similar locations of reduction and enhancement between the
liquid and graupel PSDs.

Figure 4-7: Time series and dust case minus Clean case difference plots of: ice/snow bin
particle size distribution (PSD; top row), liquid bin PSD (middle row), and graupel bin PSD
(Bottom row); averaged over the convective regime in the temperature range of -5°C to -38°C.
Contours represent \( \log_{10} \) values of bin population. Columns: Clean (a,e,i), D.12 - Clean (b,f,j),
D1.2 - Clean (c,g,k), and D12 - Clean (d,h,l) cases.
4.2.5 Effects on Convective Regime Vertical Motion and Latent Heat

The formation of smaller and more numerous cloud ice particles in the heterogeneous nucleation regime results in increased latent heat release in the convective core between -5°C and -38°C. This is due to both the diffusional growth of frozen particles and latent heat released by the phase change occurring during riming. Diffusional growth is the source of the majority of latent heat release and may consume much of the updraft’s available water vapor. Increased latent heat release invigorates convective updrafts compared to the clean case. Figure 4-8a and Figure 4-8d describe the time evolution of convective regime averaged updraft and downdraft velocity. Figure 4-8b and Figure 4-8e (Figure 8c and Figure 8f) describe the average latent heat (water vapor mixing ratio) at temperatures <0°C within the updrafts and downdrafts, respectively. As IN concentration is increased, average convective updraft intensity is progressively increased between hour 10 and 20. Likewise, updraft latent heat is increased and updraft water vapor content is reduced. This is consistent with increased diffusional growth of the more numerous particles that form in the dust cases. Increasing IN in the dust cases results in higher convective core top heights from model hour 6 to about model hour 20. During the transition to deep convection between hour 6 and hour 12 the increase in core top height is fairly linear for increasing IN concentrations. The time averaged convective core height (cloud tops <0°C) between hour 6 and 12 are: Clean (8.91 km); D.12 (8.93 km); D1.2 (9.28 km); D12 (9.34 km). Despite the invigorated updrafts occurring throughout the hour 6 to hour 20 time period, the core cloud top height is affected by the larger ice/snow particle sizes that form between hour 12 and hour 20 (Figure 4-7). The average convective core height (cloud tops <0°C) between hour 12 and 20 are: Clean (12.1 km); D.12 (12.25 km); D1.2 (12.04 km); D12 (12.61 km). Note that the average core height in the D1.2 case are lower than the Clean case during this time period due to
the presence of more large and fewer small sized particles (Figure 4-8c) as a result of the IN depletion described in Figure 2. This limits the number of particles that remain aloft in the D1.2 case, due to faster sedimentation rates of the large particles. Stronger downdrafts occurring between hour 10 and 20 results in increased evaporation of the more numerous particles in the dust cases. This consumes latent heat and increases water vapor content within the convective downdrafts (Figure 4-8e; Figure 4-8f).

**Figure 4-8:** Top row: time series of average convective updraft intensity. Time series of average latent heat (K/hr) within convective updrafts (<0°C). Time series of average water vapor content (g/kg) within convective updrafts (<0°C). Bottom row: as top row, averaged over convective downdrafts. Colors represent Clean (black), D.12 (blue), D1.2 (red), and D12 (green) cases, respectively.
4.2.6 Summary of Convective Regime Changes

In summary, increasing IN concentration in the dust cases results in increased ice formation and growth within the heterogeneous nucleation regime between -5°C and -38°C. Partitioning of available water vapor over more numerous particles shifts the PSD of cloud ice crystals to smaller sizes, which grow more slowly. The diffusional growth of these particles increases latent heat release in the heterogeneous nucleation regime and invigorates convective updrafts. Homogeneous ice formation is reduced due to fewer liquid drops crossing the -38°C threshold as well as reduced peak supersaturation due to ice growth within the heterogeneous regime. Despite reduced homogeneous ice formation, invigorated updrafts result in higher convective core cloud tops compared to the Clean case.

4.2.7 Effects on Stratiform Regime Ice/Snow PSD

The macrophysical and microphysical properties of the stratiform and anvil cloud regime are a direct consequence of cloud and precipitation formation processes initiated within the convective core. Invigorated updrafts in the dust cases carry a greater number of both large and small particles to the level of divergence. These particles are then transported by the wind field into the milder updrafts of the stratiform regime. The large particles quickly sediment out and the smaller particles remain aloft. Figure 4-9 describes the stratiform ice/snow bin distribution as Figure 4-7 described the convective ice/snow bin distribution. Between hour 6 and hour 12 in the dust cases, the initial burst of ice formed by heterogeneous nucleation in the core is transported into the stratiform regime in conjunction with local ice formation. This results in increased bin populations over much of the ice/snow PSD. After hour 12 until about hour 26, the formation of small ice particles is reduced due to the depletion of IN by ice formation earlier in the simulation.
Snow particles formed in the convective core grow to larger sizes during transport into the stratiform regime. This increases the relative bin populations at sizes between 1900um and 20000um compared to the convective regime. These large particles efficiently capture smaller particles, resulting in the greater reduction of smaller sized particles in the stratiform PSD compared to the convective regime (Figure 4-8c; Figure 4-9c). When IN concentrations are not as significantly depleted, such as in the D12 case, heterogeneous nucleation produces additional small ice crystals throughout the hour 12 to hour 26 period (Figure 4-9d). While larger sized particles continue to form in the D12 case, the location of the most significant enhancement to bin population shifts to smaller sizes compared to the D1.2 case. This is a result of competition between the more numerous particles in the D12 case during collection processes. This is visible when comparing the differences between the D1.2 and D12 case precipitation mass fractions (Figure 4-2b).
Figure 4-9: Time series and dust case minus Clean case difference plots of: ice/snow bin particle size distribution (PSD); averaged over the stratiform regime in the temperature range of -5°C to -38°C. Contours represent log₁₀ values of bin population.

4.2.8 Effects on Vertical Cloud Fraction

Many hydrometeors in the stratiform and anvil cloud regime were initially formed in the convective core and were transported into the stratiform/anvil regime by upper level winds. However, increasing IN concentration in the dust cases also results in increased heterogeneous ice formation within the stratiform regime itself (Figure 4-5b) which affects cloudiness. Figure
4-10a describes the total occurrence number of cloudy data points at each vertical level for the entire simulation. Figure 4-10(b-d) describe the horizontal fraction at each vertical level of convective (10b), stratiform (10c) and anvil (10d) clouds to total clouds, respectively. The vertical distribution of convective cloud fraction increases between -5°C and -38°C as IN concentration is increased in the dust cases. At temperatures below -38°C, such as above 12km, the convective fraction is significantly increased in the D1.2 and D12 cases compared to the Clean and D.12 cases. The invigorated updrafts in the former cases increase the frequency of overshooting convective cores. Stratiform cloud fraction is increased over the majority of the vertical range due to both increased transport from the convective core and increased heterogeneous ice formation within the stratiform regime itself. However the anvil cloud is significantly affected by changes in hydrometeor PSDs. A small IN concentration in the dust layer (D.12) results in a similar or slightly greater anvil cloud fraction compared to the Clean case. In the D.12 case, due to the limited supply of IN, the formation of large ice particles is not significantly increased compared to the small ice particles that form. The small ice particles are transported greater distances in the updrafts and sediment out slowly. Therefore the D.12 case possesses a higher and broader cloud distribution compared to the Clean case, as indicated by the increased cloud occurrence (4-10a) and reduced convective fraction (4-10d) above 11km. In the D1.2 case, some ice particles are transported from the core and more ice particles are formed locally through heterogeneous formation processes with available IN in the stratiform regime. Some of the particles grow by collection processes to large sizes (Figure 4-9c). These large particles sediment out quickly in the weaker updrafts of the stratiform regime and therefore are not transported into the anvil cloud regime. In conjunction with reduced homogeneous ice formation, this results in fewer particles forming locally within and/or being transported into the
anvil regime. Therefore the anvil cloud distribution in the D1.2 case is lower and narrower compared to the Clean case and other dust cases. The D12 case is affected by both the formation of more numerous ice large particles (compared to the D.12 case) and more numerous small ice particles (compared to the D1.2 case). The strong updraft intensities in the D12 case transport significant condensate mass into the stratiform regime. The large particles that form in the D12 case sediment out quickly, but the small ice particles remain near the cloud tops. This results in an anvil cloud distribution that is lower and less broad compared to the Clean and D.12 cases, but is higher and wider than the D1.2 case. However, after hour 20 (Figure 4-9d) the ice particles in the D12 case grow to large sizes and sediment out. This results in the lower stratiform/anvil cloud top height from hour 20 until the end of the simulation compared in the D12 case to the D1.2 case noted in Figure 4-3.
Figure 4-10: (a) Occurrence frequency of cloudy data points over total simulation time for each vertical level. Remaining figures: horizontal fraction of convective (b), stratiform (c) and anvil cloud to total cloud occurrence at each vertical level. Colors represent Clean (black), D.12 (blue), D1.2 (red), and D12 (green) cases, respectively.

4.3 Conclusions

The MCS occurring on 08 March 2004 in the tropical eastern Atlantic, first described in Min et al. (2009) was simulated using the WRF model with a spectral-bin microphysical scheme. Ice nucleation parameters within the SBM were updated to connect heterogeneous and
homogeneous ice formation with IN to investigate the effects of dust acting as IN. In the first of a two part study, we present the effects of IN activation on ice formation processes and the eventual effects on the large-scale cloud fields. The hypothesis of Min et al. (2009) suggested that dust forming ice at heterogeneous temperature ranges (-5°C to -38°C) results in changes to precipitation formation processes and ice particle size distributions shifting to smaller sizes in the heterogeneous nucleation regime. Lower stratiform/anvil cloud top heights were reported (Min and Li, 2010), despite the presence of more numerous deep clouds with large IWP which suggests that convective invigoration (increased latent heat release; stronger updrafts) is occurring.

Increasing IN concentration in the dust case simulations results in the formation of a greater number of ice particles in the convective core between -5°C and -38°C compared to the Clean case (Figure 4-6). The partitioning of available water vapor over the greater number of particles results in smaller ice crystal and graupel sizes in the dust cases (Figure 4-7). The ice particles grow more slowly due to the increased competition between individual particles for available water vapor (Figure 4-8). Latent heat release in the heterogeneous nucleation regime is increased in the dust cases due to diffusional growth and liquid-to-ice phase changes during riming of the smaller, more numerous particles. Convective updrafts are invigorated (Figure 4-2; Figure 4-8), resulting in increased overshooting and higher convective core top heights. The increased downdraft velocity and more numerous small particles result in increased evaporation and latent cooling (Figure 4-8).

Particle growth resulting from collection processes is also reduced, due to the lower collision efficiency of the smaller particle sizes in the dust cases. Therefore precipitation formation is shifted to colder temperatures (higher altitudes) within the heterogeneous nucleation
regime. This is visible when comparing the vertical profiles of TWC, precipitation fraction, and vertical rain rates of the D1.2 and D12 cases in Figure 4-2. More numerous ice crystals in the dust cases initiate the precipitation forming processes of aggregation and riming earlier in the simulation and at warmer temperatures compared to the Clean case (Chapter 5). When available IN concentration in the dust cases is depleted, the formation of new ice crystals in the heterogeneous nucleation regime is limited. Collection processes remove small ice crystals formed earlier in the simulation and increase the formation of large ice/snow particles in both the convective (Figure 4-7) and stratiform (Figure 4-9) regimes. This is most visible in the D1.2 case between hour 12 and hour 18. The large snow particles fall out quickly after being transported into the milder updrafts of the stratiform regime. When few small particles remain aloft, due to reduced homogeneous ice formation and/or increased particle collection, stratiform/anvil cloud top heights will be lower over the majority of the simulation, as in the D1.2 and D12 cases (Figure 4-3; Figure 4-10). When small particles are relatively more numerous, such as in the D.12 case, stratiform/anvil cloud top heights will be higher than the Clean case (Figure 4-3; Figure 4-10). The small particles in the D.12 case are transported to higher altitudes in the convective updrafts and remain aloft for longer times.

The dust affected reflectivity profiles reported in Min et al. (2009) and further expanded upon in Li and Min (2010) have been partially reproduced in our simulations (Figure 4-4). Increasing IN concentration results in reduced convective reflectivity at temperatures above 0°C and increased stratiform reflectivity at temperatures below 0°C, consistent with observations. However, convective reflectivity below 0°C and stratiform reflectivity above 0°C were both increased in the dust cases compared to the Clean case. Observations indicate that reflectivity is reduced over these temperature ranges in the respective cloud regimes. Radar reflectivity is
sensitive to changes in the number concentration and PSD of precipitation particles due to the large sizes of such particles. Therefore, additional analysis of the precipitation formation processes must be undertaken to better understand the feedbacks between dynamics and microphysics in DCC forming under high IN concentrations. Chapter 5 will focus on the sensitivity of convective and stratiform precipitation formation processes to different IN number concentrations and the resulting effects on radar reflectivity.
Chapter 5

Changes to Precipitation Processes

Abstract

To better understand the impacts of dust aerosols on deep convective cloud (DCC) systems, a case study in the tropical eastern Atlantic was investigated using the Weather Research and Forecasting (WRF) model coupled with a Spectral Bin Microphysics (SBM) model as a two part study. In Part two, we focus on the changes to microphysical processes that impact precipitation formation. Total surface precipitation accumulation is reduced proportionally as IN concentration is increased, due to less efficient graupel formation reducing convective rain rates. Stratiform precipitation accumulation is increased due to greater snow formation and growth, but does not counteract the reduced convective accumulation. Competition between the more numerous particles for available water vapor during diffusional growth and available smaller crystals/drops during collection reduces particle growth rates and shifts precipitation formation to higher altitudes in the heterogeneous nucleation regime. Rimming efficiency in the dust cases is reduced due to smaller cloud ice crystals, resulting in smaller graupel sizes overall. Ice particle aggregation occurs earlier in the simulation and over a wider range of temperatures in the dust cases, which increases snow formation in the heterogeneous nucleation regime. Radar reflectivity values are increased in the dust cases at temperatures below 0°C in both the convective and stratiform regimes due to more large snow particles. More numerous small ice/graupel particles that form in the heterogeneous nucleation regime in the dust cases melt and
reduce reflectivity values in the convective core near the surface, consistent with case study observations.

5.1 Introduction

Mesoscale convective systems (MCSs) have a large impact on regional cloud cover, radiative transfer, and atmospheric circulations. They are also significant sources of precipitation especially in the tropics (Arakawa, 2004; Solomon et al., 2007). Observational and modeling studies of DCC have shown different results relating to the effect of aerosol on convection and precipitation, indicating that aerosol may either enhance or suppress convection and precipitation depending on aerosol concentration and environmental conditions (Khain and Pokrovsky, 2004 Khain et al., 2004, 2005, 2008; van den Heever et al., 2006; Fan et al., 2007b; Min et al., 2009; Min and Li, 2010; Li and Min, 2010; Min et al., 2014; Altaratz et al., 2014). Chapter 5 describes the effects of different IN concentrations on processes that affect precipitation formation in deep convective clouds to compliments the results presented in Chapter 4. In addition, a sensitivity study testing the effects of moisture content in the model’s dust layer is described below to provide further insight into the differences between observed changes to radar reflectivity and the radar reflectivity simulated by our model in this study.

Polarimetric radar measurements hold great promise in expanding the knowledge of microphysical processes occurring within deep convective clouds where in situ measurements are not feasible. Dual-polarized radar has the potential to significantly impact cloud modeling by improving microphysical parameterizations or by directly assimilating measurements into numerical models (Ryzhkov et al., 2011). For example, algorithms can be used to realistically discern hydrometeor type from polarimetric radar observations (Thompson et al., 2014) to
evaluate microphysical processes within numerical models. However, radar reflectivity is sensitive to particle qualities such as particle number, size, shape, density, orientation, and water content. For this reason, detailed numerical models that are capable of tracking such qualities are in turn valuable for evaluating and interpreting radar measurements when in situ verification is not possible (Ryzhkov et al., 2011). The use of detailed bin microphysics allows for explicit calculation of microphysical processes that affect cloud and precipitation formation and growth. In addition, the bin PSDs can be directly converted into radar reflectivity values which can then be compared with observations.

5.2 Results

5.2.1 Changes within Individual DCC

Increasing IN concentration in the dust cases results in greater ice formation and growth within the heterogeneous nucleation regime. This affects homogeneous ice formation by reducing the number of liquid drops that reach the -38°C threshold and also by reduced peak supersaturation values due to the growth of more numerous ice particles within the heterogeneous nucleation regime. Figures 5-1 and 5-2 depict the vertical cross-section of a specific convective core and its associated stratiform/anvil cloud at a single model time step (hour 15) from the Clean and D1.2 cases. The cross-section slices are not identically located in the two cases due to small differences in the spatial evolution of the system, but are less than 4 grid points apart. In both cases, the slices are similarly located within their respective cloud system and are at similar stages of evolution. The slices are averaged zonally over 9km to further reduce the effects of spatial variations. The Black and dashed blue lines (Figure 5-1; Figure 5-2) depict updrafts (> 1m/s) and downdrafts (< -0.1 m/s). The grey dashed line (Figure 5-1; Figure 5-
2) depicts the threshold value of cloudiness suggested by Fan et al. (2013) and shows the change to cloud geometry directly. The Clean case (Figure 1a) shows the classic DCC cloud structure of convective core and associated stratiform region transitioning into the anvil. The D1.2 case also possesses a similar cloud structure, but with a far smaller anvil cloud, which is a result of the changes to the partition between homogeneous and heterogeneous ice formation in the D1.2 case. Cloud formation is increased in the heterogeneous nucleation regime (Figure 5-1d) compared to the Clean case (Figure 5-1a). Liquid drops that would otherwise freeze at temperatures colder than -38°C are converted to ice at warmer temperatures due to increased riming and/or immersion/contact nucleation. In addition, increased ice formation and growth within the heterogeneous nucleation regime reduces peak supersaturation values at colder temperatures, limiting ice formation in the homogeneous regime. Therefore, fewer particles form within and/or are transported into the anvil regime which limits its extent compared to the Clean case.

The first column of Figure 5-1 describes total water content (TWC), while columns 2 and 3 describe rain rate and radar reflectivity, respectively. TWC is increased in the dust case (Figure 5-1d) at temperatures below 0°C compared to the Clean case (Figure 5-1a). The higher TWC in the heterogeneous nucleation regime is accompanied by a correspondingly larger area of strong vertical motion. This supports the evidence of convective invigoration due to increased latent heat release in the dust-affected deep, high-IWP clouds reported by Min and Li (2010). Stronger updrafts in the dust cases supply sufficient water vapor to support the formation and growth of more numerous particles in the heterogeneous nucleation regime and can transport a greater number of large particles to higher altitudes in the convective core and into the adjoining stratiform regime. These large particles contribute to the higher rain rate values noted in the D1.2
case (Figure 5-1e) compared to the Clean case. The increased rain rates at temperatures below 0°C also correspond to the increased radar reflectivity values in the stratiform regime from the convective core almost to the anvil regime near the equator (Figure 5-1f). Figure 2 describes the effective radii (Re; 1.e2 um) of rain drops, graupel, and snow particles in columns 1-3, respectively. Rain drop radii are significantly decreased in the heterogeneous nucleation regime due to large sized drops freezing by immersion/contact nucleation or by collisions with ice particles (riming; Figure 5-5) leaving smaller drops unfrozen. Graupel and snow radii are both decreased at temperatures below 0°C. This reduction is most pronounced within the convective core where competition between more numerous small particles during collision-collection reduces growth rates (Figure 5-5; Figure 5-7). At temperatures above 0°C, graupel radii is increased in the dust cases due to immersion freezing of large rain drops into graupel within the heterogeneous nucleation regime and then falling into warmer temperatures. At temperatures below -38°C and in the anvil cloud regime, graupel and snow radii are increased compared to the Clean case. This is due to the stronger outflow in the dust cases from the convective core, which transports large precipitation particles greater distances before they sediment out of the cloud. In addition, precipitation formation is shifted to colder temperatures (higher altitudes) in the heterogeneous nucleation regime which increases the number of large particles forming near the cloud tops.
Figure 5-1: Zonally averaged longitude slice plot of similar DCC structures within the Clean (row 1) and D1.2 (row 2) cases. Shaded colors: total water content (TWC; column 1), vertical rain rate (column 2), and radar reflectivity (column 3); Line contours, all columns: vertical motion (solid black >1m/s; dashed blue <-0.1m/s); cloudiness threshold (dashed grey, >1e-6 kg/kg).
Figure 5-2: Slice plots representing same DCC as in Figure 1 for Clean (row 1) and D1.2 (row 2) cases. Shaded colors: rain drop effective radii (Re; column 1), graupel Re (column 2), and snow Re (column 3); Line contours, all columns: vertical motion (solid black >1 m/s; dashed blue <-0.1 m/s); cloudiness threshold (dashed grey, >1e-6 kg/kg).

5.2.2 Radar Reflectivity Dry Layer Sensitivity

With advances in observing technology, cloud and precipitation radars are used extensively for studying cloud and precipitation formation and microphysical-dynamical interactions. Min et al. (2009) used contoured frequency by altitude (CFAD) plots to describe the observed changes to convective and stratiform radar reflectivity in the dust-free and dusty regions. They noted that radar reflectivity at temperatures above 0°C was reduced in the dusty region for both the convective and stratiform regimes. At temperatures below 0°C, convective
reflectivity was reduced in the dusty regions while stratiform reflectivity was increased. To compare the observations of Min et al. (2009) with our results, we have recreated similar CFAD plots using model derived radar reflectivity. Figure 5-3 describes the radar reflectivity CFADs of the convective and stratiform regimes for the Clean case and dust case minus Clean case difference plots. The black contour line in the difference plots (columns 2-4) represents the Clean case’s 2% contour value outline for the respective convective or stratiform CFAD plot in column 1.

As IN concentration is increased, convective reflectivity at temperatures above 0°C is decreased. Likewise, stratiform reflectivity at temperatures below 0°C is increased in the dust cases. These changes suggest that increased heterogeneous ice formation is significantly affecting the formation of precipitation sized particles, consistent with the hypothesis of Min et al. (2009). We note that convective reflectivity at temperatures below 0°C and stratiform reflectivity at temperatures above 0°C are both increased in the dust cases compared to the Clean case. This differs from the reduced reflectivity values reported in Min et al. (2009) and Li and Min (2010) for these locations. These differences can be partially explained by greater water vapor content within the dust layer in the model simulations compared to the observed SAL.

Measurements from AIRS/AMSU/HSB indicate that the relative humidity in the dust layer is about 20% drier than the surrounding air. While a dry air layer is present in the WRF’s initial and boundary conditions, the model slightly overestimates precipitable water compared to observations (Gibbons et al., 2017). To examine the impacts of dust layer moisture content on our case study, we have conducted additional numerical simulations based on the D1.2 case. The dust layers within these test cases feature relative humidity values that are 5% drier than the original D1.2 case. The first case (Dry5init) reduces the water vapor content in the dust layer
over the entire 4\textsuperscript{th} domain at model start-up time. The boundary conditions entering the 4\textsuperscript{th} domain are unchanged from the original D1.2 case. The second case (Dry5bound, not shown) reduces water vapor content at the boundaries of the 4\textsuperscript{th} domain for the duration of the simulation with no changes made to the model’s initialized moisture content. Figure 5-4 describes the convective and stratiform CFADs of the D1.2 and Dry5init cases. The first and second columns describe the D1.2 CFAD and the D1.2 minus Clean case difference plots, respectively. The third column describes the Dry5init minus Clean case difference plots. Reduced moisture content in the Dry5init case weakens convective cloud formation, which decreases convective reflectivity overall at temperatures below 0°C and shifts reflectivity to lower values at temperatures above 0°C. Reflectivity in the stratiform regime is still increased compared to the Clean case at temperatures below 0°C, but is also shifted to lower values at temperatures above freezing. These changes are very similar to the observed changes of convective and stratiform reflectivity described by Min et al. (2009) and Li and Min (2010). The Dry5bound case results in similar changes as those described by the Dry5init, although with greater reductions in the convective regime and smaller increases in the stratiform regime, although with greater reductions in the convective regime and smaller increases in the stratiform regime as a result of the drier boundary air transported into the 4\textsuperscript{th} domain for the duration of the simulation.
Figure 5-3: Contoured frequency by altitude diagrams (CFAD) of model simulated convective (top row) and stratiform (bottom row) radar reflectivity. Columns: Clean case, D1.2-Clean, D1.2-Clean, D12-Clean difference plots, respectively. Black contour line in difference plots represents the Clean case 2% contour value.
5.2.3 Effects on Vertical Profiles of Hydrometeor PSDs

The initial PSD of the ice particles formed by heterogeneous freezing is significantly affected by the dominant freezing mechanism. The deposition freezing mechanism exclusively forms small ice crystals while the immersion and contact mechanisms freeze liquid drops starting from the largest sizes to the smallest. This affects the processes of vapor growth and particle collection growth by increasing competition for water vapor (due to a greater number of
particles) and reducing collection efficiency (due to smaller particle sizes), respectively. By limiting growth in this way, the hydrometeor PSDs will be affected both above and below the freezing level as particles experience different dynamical and thermodynamical regimes. The use of bin microphysics allows us to consider subtle changes within the PSD that may result from nucleation, growth, evaporation and collection that would otherwise be obscured if the PSD shape was prescribed as in many bulk models. Figure 5-5 describes the time averaged vertical profiles of convective regime PSD bin populations. The vertical profile averages of the Clean and dust cases are depicted by color contours, while the dust minus Clean differences (columns 2-4) are represented by line contours (Black: +0.5, +1.0, +2.0; Red: -0.5, -1.0, -2.0). The provided radii values of the represented hydrometeor species are taken from the pre-calculated bin radii values used by the model, which are based on assumed particle densities and the mass doubling relationship between the individual bins. Contour values represent log values of bin population. The difference plots likewise describe the relative change of these log values, representing Log(Dust/Clean) values.

Ice formation in the dust cases results in fairly consistent changes to the ice/snow PSD as IN number is increased. Heterogeneous ice formation increases the ice/snow PSD bin concentrations between -5°C and -38°C overall with the location of maximum enhancement shifting towards smaller cloud ice sizes. This is due to water vapor being partitioned over more particles during initial formation and from greater diffusional growth competition between these more numerous ice crystals. At temperatures above 0°C, the greater ice content formed in the heterogeneous nucleation regime and transported down results in a similar peak enhancement location at smaller particle sizes. The bin population of large snow particle sizes is also increased in the dust cases (Figure 5-5) due to earlier and more frequent ice crystal aggregation (Figure 5-
6) resulting from increased ice particle number concentrations in the heterogeneous nucleation regime. At temperatures colder than -38°C the ice/snow PSD is narrowed compared to the Clean case, with both the largest and smallest sizes reduced. Collection processes occurring at warmer temperatures in the dust cases reduce the number of small particles transported into the homogeneous freezing regime from warmer temperatures. More numerous particles in the heterogeneous nucleation regime compete for available vapor and smaller particles for collection, which limits the growth of larger particles sizes. Liquid drop PSD is reduced over the majority the size range due to the conversion of liquid drops to ice by riming or immersion/contact freezing in the heterogeneous nucleation regime. The population of larger drops is increased in the dust cases due to the greater re-circulation of recently melted large ice particles in the stronger updrafts. Larger drops may also be lifted into the heterogeneous nucleation regime from warmer temperatures by these stronger updrafts. The graupel PSD is affected by both increased freezing of large rain drops by immersion or contact freezing and by increased riming of more numerous but smaller ice crystals. The majority of graupel are formed by the latter process, which results in the graupel PSD shifting to smaller sizes overall as IN concentrations are increased.

We noted in Chapter 4 (Gibbons et al., 2017) that IN depletion has a significant effect on the PSD of ice and snow in the D1.2 case compared to the other two dust cases. In the D1.2 case more significant ice formation occurs in the heterogeneous nucleation regime compared to the D.12 and Clean cases as a result of the larger IN concentration. The greater ice formation initiates particle collection processes earlier in the simulation and over a wider range of temperatures compared to the Clean and D.12 cases. When the simulation’s available IN is depleted in the D1.2 case after about hour 12, few new ice crystals form in the heterogeneous
nucleation regime and ice crystals formed earlier in the simulation are collected into snow particles. These snow particles sediment out quickly after being transferred into the stratiform regime. In addition, relatively fewer particles remain aloft at temperatures below -38°C due to the reduced formation of ice particles in the homogeneous regime as noted previously. This reduces reflectivity values above 9km in the D1.2 case compared to the other dust cases (Figure 5-3). While IN depletion occurs in both the Clean and the D.12 case, heterogeneous ice formation is not as significant as in the D1.2 case. Small ice particles still form by homogeneous drop freezing in these cases without being depleted by aggregation into larger snow particles. In the D12 case, IN depletion is significantly less severe than the D1.2 case due to the greater total IN number. This allows small ice crystals to continue forming in the heterogeneous nucleation regime despite significantly increased snow formation compared to the D1.2 case (Figure 5-6).
Figure 5-5: First column: Clean case time averaged vertical profile of convective particle size distribution (PSD) bin populations (color contours; log\textsubscript{10}) for ice/snow (row 1), cloud/rain (row 2), and graupel (row 3), respectively. Second, third, and fourth columns: time averaged convective PSDs for dust cases (color contours) and dust case minus Clean case differences (line contours). Black line contour log\textsubscript{10} values: +0.5; +1.0; +2.0. Red line contours log\textsubscript{10} values: -0.5; -1.0; -2.0.

5.2.4 Effects on Collection Processes and Particle Fall Rates

Particle collection processes are the primary source of precipitation formation due to the more rapid accumulation of mass compared to purely diffusional growth. In liquid clouds,
collision-coalescence processes allow cloud drops to collect into rain drops. In ice and mixed phase clouds, ice-ice (aggregation) collisions and ice-liquid (riming) collisions become more frequent as total ice number increases. Figure 5-6 describes the changes to aggregation (row 1), riming (row 2), and drop autoconversion (row 3) in the convective regime with respect to time for the Clean and dust cases. Drop autoconversion rate (1.e-4 kg m^-1 s^-1) tracks the formation of rain drops from cloud drops by collision-coalescence processes. Aggregate number (kg^-1) tracks the change of ice particles before and after aggregation occurs and is more negative for a more efficient process. Rimming rate (g kg^-1 s^-1) tracks the liquid mass converted to graupel through the riming process and, again, is more negative for a more efficient process. These two processes are also affected by the relative availability of liquid and ice content within the cloud. As riming can only occur where ice and liquid particles coexist, this limits the most significant riming to the convective core below the cloud’s glaciation level. Likewise the drop collision-coalescence processes are reduced in the heterogeneous nucleation regime in the dust cases due to the conversion of liquid content into ice at temperatures below 0°C. In the stratiform regime very little liquid content is transported from the convective core due to the majority of freezing occurring in the core itself. Therefore, ice-ice particle interactions are the most common in the stratiform regime and snow is the predominant precipitation particle type (Stith et al., 2002; Heymsfield et al., 2002; Lawson et al., 2010, Gallagher et al., 2012).

In the Clean case the majority of ice forms homogeneously, which limits significant ice-ice particle interactions in the heterogeneous nucleation regime until a significant number of ice particles have fallen down from the homogeneous freezing regime. The small addition of IN in the D.12 case form a sufficient number of ice particles to increase aggregation activity near the 0°C freezing level, but a noticeable gap remains due to the greater percentage of homogeneous
freezing compared to the other dust cases. Increasing the IN concentration further results in maximum values near 0°C and decreasing upwards to colder temperatures. The significantly larger values of aggregation number in the D12 case compared to the other cases (Figure 5-6a to Figure 5-6d) is a result of the greater number of ice crystals forming at warmer temperatures where particle “sticking” efficiency is higher (Hallgren and Hosler, 1960). While aggregation is the primary precipitation process in the stratiform regime, the aggregation numbers in this regime are smaller than in the convective regime. This is a result of the significantly greater number of ice crystals that form initially in the core and are subsequently collected into snow particles before being transported into the stratiform regime.

The effect of IN concentration on the efficiency of riming is tied into both the size and number of ice particles and the overall availability of liquid water drops. The larger midlevel liquid water content in the Clean case results in efficient riming despite the lower midlevel ice number compared to the dust cases. Increased ice formation in the heterogeneous nucleation regime increases riming rates near the freezing level. This is due to the greater total number of ice particles and the significant presence of liquid water content near the melting level. Above 6km in the convective regime, where ice formation becomes significant in the dust cases, riming rates become progressively lower as IN concentration is increased. The smaller sizes of ice particles forming in these locations reduce the collision efficiency between ice particles and liquid drops. The reduced number of available liquid drops in the dust cases also affects riming rates by decreasing the depth of the mixed-phase environment in which riming may occur. Changes to drop autoconversion rates are similarly affected by changes to drop number concentration and PSD. While the current case studies do not allow for dust particles to activate as both CCN and IN, changes to collision-coalescence processes result from changes to ice
formation and the subsequent impact on liquid water mass both above and below the freezing level. In general the addition of IN to the dust cases results in lower liquid water content in the heterogeneous nucleation regime due to riming and immersion/contact drop freezing which limits the opportunities for collision-coalescence to occur. At altitudes below 6km, collision-coalescence rates are affected by the number and PSD of ice particles that melt after falling into above freezing temperatures. We note that higher autoconversion numbers occur at temperatures slightly above 0°C between hour 15 and 20 in the D1.2 (row 3c) and D12 (row 3d) cases. These increases are also visible in the vertical rain rates at these temperatures (Figure 5-9).
Figure 5-6: Time series of convective averaged aggregate number (row 1), riming rate (row 2), and drop autoconversion (collision-coalescence) number. Columns: Clean, D.12, D1.2, and D12 cases, respectively.

Changes in fall rates of precipitation sized particles such as snow and graupel will affect eventual surface precipitation accumulation by changing below-cloud particle residence times and subsequent evaporation. Figure 5-7 provides the convective (row 1) and stratiform (row 2) averaged particle fall rates (cm s⁻¹) for snow (column 1 and 2), and graupel (column 3 and 4) particles, averaged over the total simulation time. Dust case minus Clean case differences for convective and stratiform data are provided in column 2 and row 4, respectively. Particle fall rate
is determined by combining calculated particle terminal velocities (positive downwards, Khain and Sednev, 1995) with vertical velocity (positive upwards). In Figure 5-7 (column 1 and 3), positive numbers indicate motion towards the surface. Snow particles tend to grow larger by aggregation processes at warmer temperatures due to greater “stickiness” (Hallgren and Hosler, 1960), which results in fall rates generally increasing towards the surface. This is clearly visible in the stratiform regime (Figure 5-7e) due to mild vertical velocities. In the convective regime, smaller particle sizes and stronger updrafts in the dust cases reduce fall rates in the heterogeneous nucleation regime as IN concentration is increased. Snow particles affect radar reflectivity values in the convective regime due to the relatively large diameter of snow particles for a given particle mass compared to rain drops or graupel. While snow particles shift to smaller sizes on average in the dust cases, reduced fall rates also allow more particles to remain suspended in the cloud for a longer time. The increased formation of snow in the dust cases is significant enough to cancel out the reduced reflectivity values that result from the shift to smaller sized rain and graupel particles in the heterogeneous nucleation regime (Figure 5-2; Figure 5-5). At temperatures below -38°C, fall rates are increased in the D1.2 and D12 cases due to both the increased vertical transport of large particles in the stronger updrafts and stronger downdrafts in general. While average snow particle fall rate is higher in the D12 case compared to the D1.2 case at these altitudes, particles are also more numerous in the D12 case. Hence, radar reflectivity above 9km (Figure 5-3) is increased in the D12 case but is minimally affected (convective regime) or is decreased (stratiform regime) in the D1.2 case. Larger sizes of graupel and snow settle out more quickly overall and tend to accumulate around the melting level, increasing radar reflectivity values between 2km and 5km (Figure 5-3).
Figure 5-7: Particle fall rates (cm s\(^{-1}\)) for snow (column 1 and 2), and graupel (column 3 and 4) particles time averaged over total simulation time. Row 1 and row 2 depict convective and stratiform regime averages, respectively. Column 1 and column 3 depict average fall rates. Column 2 and column 4 represent the dust case minus Clean case differences of average fall rate. Fall rate is calculated by combining particle terminal velocity (positive downwards) with vertical motion (positive upwards).

5.2.5 Precipitation Water Content and Surface Accumulation

Increasing IN concentrations in the dust cases results in more numerous ice particles growing from initially smaller sizes (Figure 5-5). These particles compete with each other for available water vapor, which reduces the total growth of individual particles. This yields significant changes to the ratio of cloud and precipitation particles in the convective core,
shifting the formation of precipitation to higher altitudes and increasing latent heat release within the heterogeneous nucleation regime due to increased total ice mass and reduced sedimentation rates of smaller particles. Figure 5-8 describes the water content of snow (row 1), graupel (row 2), and rain (row 3) averaged with respect to temperature and vertical motion. Particle collisions are the primary mechanism by which precipitation sized particles form and are affected by changes to particle number concentration and PSD. The shift of ice formation from colder temperatures (<-38°C) to the heterogeneous freezing regime (-5°C to -38°C) is accompanied by greater snow water content as the more abundant ice crystals aggregate over a larger temperature range compared to the Clean case (Figure 5-6). However, when the cloud ice content is sufficiently large, such as in the D12 case, snow formation is slightly hindered due to greater collection competition between the individual particles. This is visible in the smaller area of maximum snow water content in row 1d compared to row 1c and the reduced updraft snow water content at temperatures warmer than -10°C. The formation of graupel shows similar signs of competition between particles, as the freezing level maximum is only increased from the Clean case value in the D.12 case and is progressively reduced at higher IN numbers. Graupel number is uniformly increased as IN number is increased (Chapter4; Gibbons et al., 2017) but riming rates are decreased (Figure 5-6), which indicates the individual particles are less efficient at gaining mass when riming due to their reduced sizes (Figure 5-5). Graupel water content is higher in the dust cases for temperatures colder than -20°C, due to both the greater number of large frozen drops and more frequent riming of more numerous small ice formed at warmer temperatures in the heterogeneous nucleation regime and transported vertically in the stronger updrafts.
Figure 5-8: Hydrometeor water content averaged with respect to temperature and vertical motion regime for: snow (row 1), graupel (row 2) and rain (row 3) hydrometeors, respectively.

Columns: Clean, D.12, D.1.2, and D12 cases, respectively.

Increasing heterogeneous ice formation by increasing IN concentrations results in larger ice mass near the 0°C temperature level, but greater competition during growth between individual particles for water vapor and available small drops/crystals for collection shifts the formation of precipitation sized particles to higher altitudes. Smaller particles that sediment out or are transported below the melting level are more likely to evaporate below the cloud due to a
slower fall speed. These changes result in a reduced surface accumulation and enhanced rain rates above the freezing level. Figure 9 describes the accumulated surface rain rates (Figure 5-9a, Figure 5-9b) and rain rate vertical profile differences (Figure 5-9d, Figure 5-9e) for convective (column 1) and stratiform (column 2) regimes. Figure 5-9c describes the time series of total accumulated surface precipitation, while Figure 5-9f describes the total fraction of precipitation formed at each vertical level, for the Clean and dust cases. In general, the addition of IN reduces the average surface rain accumulation for the convective (Figure 5-9d) rain regime and increases it for the stratiform (Figure 5-9e) rain regime. This is due to the different effects of dust on the primary sources of precipitation in the two regimes.

Convective rain is significantly affected by changes to graupel formation, which in the dust cases is shifted towards smaller sizes (Figure 5-5). The smaller graupel sizes are a result of both the reduced formation of graupel by immersion freezing and decreased riming rates above 6km due to smaller ice sizes and lower liquid water content. In the stratiform regime precipitation is predominantly a result of snow formation. In the dust cases, snow formation is enhanced due to the increased transport of ice mass from the convective core and the warmer glaciation temperatures in the convective regime. This initiates the aggregation processes earlier in the simulation and at warmer temperatures than in the Clean case (Figure 5-6). At altitudes below 6km, collision-coalescence rates are affected by the number and PSD of the frozen particles that melt in the above-freezing temperatures. In the convective regime, increased aggregation rates (Figure 5-6) and freezing of large drops to graupel (Figure 5) result in higher autoconversion (collision-coalescence) rates in the D12 case compared to the other dust cases between 1km and the freezing level when these large particles melt. This partially counteracts the reduced rain rates between 4km and 8km resulting in near surface rain rates that slightly
exceed the D1.2 case, although final surface accumulation is still lower in the D12 case due to the greater reductions at higher altitudes.

**Figure 5-9:** (a) and (b) Time series of accumulated surface precipitation for convective and stratiform data respectively. (c) Total accumulated surface precipitation for the clean and dust cases. (d) and (e) dust case minus Clean case time averaged vertical rain rates for convective and stratiform precipitation, respectively. (f) Fraction of total precipitation formed at each vertical level. Colors represent: Clean (black), D.12 (blue), D1.2 (red), D12 (green) cases, respectively.
5.3 Conclusions

The MCS occurring on 08 March 2004 in the tropical eastern Atlantic, first described in Min et al. (2009) was simulated using the WRF model with a spectral-bin microphysical scheme. Ice nucleation parameterizations within the SBM were updated to connect heterogeneous and homogeneous ice formation with aerosols to investigate the effects of dust acting as IN. Total surface precipitation accumulation is reduced as IN concentration is increased (Figure 5-9), due to less efficient graupel formation (Figure 5-6; Figure 5-8) that reduces convective rain rates. Stratiform precipitation accumulation is increased due to greater snow formation and growth (Figure 5-6; Figure 5-8), but this does not offset the reduced convective accumulation. Instantaneous cross-section plots comparing similar DCC (Figure 5-1; Figure 5-2) indicate that adding IN in the dust cases results in greater TWC in the heterogeneous nucleation regime, smaller average precipitation particle radii, increased (reduced) rain rates at temperatures below (above) 0°C, and corresponding increases/reductions to radar reflectivity values.

Increasing IN concentration in the dust cases results in greater formation of small ice crystals in the heterogeneous nucleation regime (Chapter 4; Figure 4-5; Figure 4-8). The larger concentration of small ice crystals in the dust cases increases the conversion of liquid drops to ice mass between -5°C and -38°C, due to increased riming and/or immersion drop freezing (Figure 5-5; Figure 5-6). Homogeneous ice formation is reduced in the dust cases due to both fewer liquid drops crossing the -38°C threshold and reduced peak supersaturation values resulting from greater ice diffusional growth in the heterogeneous nucleation regime. More numerous but smaller graupel particles form in the dust cases (Chapter 4; Figure 4-2; Figure 4-5) due to the reduced riming efficiency of small ice particles (Figure 5-6) and increased competition between the individual frozen particles during riming for available liquid drops. The greater
heterogeneous ice numbers also increase ice particle aggregation in the -5°C to -38°C temperature range, leading to increased snow formation in both the convective and stratiform regimes. Growth competition between the more numerous individual particles during riming/aggregation shifts precipitation formation to higher altitudes within the heterogeneous nucleation regime. This results in changes to simulated reflectivity values (Figure 5-3) which are similar to observed effects on reflectivity (Min et al., 2009; Li and Min, 2010).

The impacts of dust as IN on model simulated reflectivity are mostly consistent with observed changes, i.e., dust cases producing smaller reflectivity values near the surface and larger values above the freezing level and most significantly in the stratiform regime (Figure 5-1; Figure 5-3). Radar reflectivity in the dust cases is affected by PSDs shifted to smaller sizes, reduced particle fall rates, and increased formation of snow particles. The contribution of graupel and rain drops to total reflectivity in the dust cases is reduced due to the shift to smaller particle sizes (Figure 5-2; Figure 5-5) and reduced drop concentrations (Figure 5-8), respectively. This decreases dust case reflectivity values at temperatures above 0°C in the convective regime (Figure 5-1, Figure 5-3). Snow particle have large radii compared to graupel and rain drops of comparable mass (Figure 5-2; Figure 5-5) and have slower fall rates (Figure 5-7). More numerous large snow particles in the dust cases result in increased reflectivity values at temperatures below 0°C (Figure 5-3), most notably in the stratiform regime where aggregation is the dominant precipitation formation process. The dust case reflectivity CFADs differed from observed reflectivity changes in the convective regime (>0°C) and in the stratiform regime (>0°C). Specifically, reflectivity in these locations is increased in the dust case simulations while observations indicate that reflectivity is reduced. Higher moisture content in the dust layer compared to the observed test cases was suggested as a possible cause of these differences.
Additional test cases based on the D1.2 case were simulated to determine the effects of reduced moisture content within the dust layer on model results. Reducing dust layer moisture content by 5% (Dry5init case) was sufficient to weaken convective cloud formation and affect the resulting reflectivity CFADs (Figure 5-4) in ways consistent with observed changes (Min et al., 2009; Li and Min, 2010). Convective reflectivity (<0°C) and stratiform reflectivity (>0°C) were both reduced compared to the Clean case. Stratiform reflectivity at temperatures below 0°C was also increased from the Clean case, indicating that microphysical changes to cloud and precipitation formation processes are similar to those in the original D1.2 case.
Chapter 6

Changes to Latent Heat and Atmospheric Circulation

6.1 Introduction

Mesoscale convective systems (MCS) have a significant impact on regional cloud cover, radiative transfer, and atmospheric circulations. They are also significant sources of precipitation especially in the tropics (Arakawa, 2004; Solomon et al., 2007). Case study observations have reported that dust aerosols have strong microphysical impacts on MCSs for similar dynamical conditions (Min et al., 2009; Min and Li, 2010; Li and Min, 2010; Min et al., 2014) which may alter precipitation vertical structures, shift the precipitation size distribution from heavy to light precipitation, and even suppress precipitation entirely. These changes in precipitation vertical structure can have a profound impact on the latent heat profiles of the atmosphere and subsequently regional and global circulations (Tao et al., 2006, 2007).

Ice cloud formation is strongly dependent on temperature, moisture saturation ratio and the strength of the convective core, the latter two of which are currently impossible to observe by satellite. Therefore the use of a detailed numerical model is crucial in untangling the effects of microphysics from dynamical and thermodynamical processes. In order to do this we have implemented a wide array of ice formation processes that cover numerous different mechanisms in the temperature range between 0 and -38 degrees Celsius as described in Chapter 3. These mechanisms directly connect the presence of dust aerosols to ice formation. In Chapter 6 we will focus in detail on the latent heat and cooling processes affected by microphysical changes in our
three dust cases and the subsequent impacts on atmospheric circulation and relative humidity in the overall 4th domain.

6.2 Results

6.2.1 Changes to Specific DCC

The majority of latent heat exchange in the atmosphere is a result of vapor-particle growth and/or evaporation/sublimation phase changes processes. Liquid-solid-liquid phase changes do not affect total hydrometeor mass and contribute only around 10% of total latent heat (Li et al., 2013; Min et al., 2013). Dust aerosols can affect latent heat processes by changing hydrometeor number concentrations and PSD as a result of increased heterogeneous ice formation and growth. Figures 6-1 to 6-3 depict the vertical cross-section of a specific convective core and its associated stratiform/anvil cloud at a single model time step (hour 15) from the Clean and D1.2 cases. The cross-section slices are not identically located in the two cases due to small differences in the spatial evolution of the system, but are less than 3 grid points apart. In both cases, the slices are similarly located within their respective cloud system and are at similar stages of evolution. The slices are averaged zonally over 9km to further reduce the effects of spatial variations. The Black and dashed blue lines (Figures 6-1 to 6-3) depict updrafts (> 1m/s) and downdrafts (< -0.1 m/s). The grey dashed line (Figures 6-1 to 6-3) depicts the threshold value of cloudiness suggested by Fan et al. (2013) and shows the change to cloud geometry directly. The color contours in Figure 6-1 describe liquid water content (LWC, Column 1), ice water content (IWC, column 2), and vertical motion (column 3), respectively. Increased heterogeneous ice formation and growth in the D1.2 case reduces LWC at temperatures below 0°C compared to the Clean case due to increased riming of frozen particles (Chapter 5; Gibbons
et al., 2017) and/or immersion drop freezing. Conversely, LWC at temperatures above 0°C is slightly increased in the D1.2 case due to the increased melting of ice particles that form in the heterogeneous nucleation regime (-5°C to -38°C). Increased heterogeneous ice formation and growth in the D1.2 case increases IWC at temperatures below 0°C in both the convective and stratiform regimes. Strong vertical motion in the convective core increases transport of frozen particles to temperatures below -38°C. However, homogeneous ice formation is reduced in the dust cases (Chapter 4 & 5; Gibbons et al., 2017) due to increased conversion of liquid drops to ice at temperatures above -38°C and reduced peak supersaturation values resulting from greater diffusional growth at warmer temperatures in the heterogeneous nucleation regime. The change in the partition between homogeneous and heterogeneous ice formation in the dust cases results in reduced IWC content in the stratiform/anvil regime at temperatures below -38°C (Figure 6-1e). Convective core updraft intensity in the D1.2 case is greater with a maximum value located at warmer temperatures (lower altitudes) compared to the Clean case, due to increased latent heat release during the growth of the more numerous ice particles (Figure 6-1e; Figure 6-2e; Chapter 4 & 5).

Figure 6-2 describes latent heat (LH) resulting from condensation/evaporation, deposition/sublimation, and freezing/melting processes in color contours, (columns 1-3, respectively). The reduced LWC at temperatures below 0°C in the D1.2 case (Figure 6-1d) results in a corresponding reduction in condensation LH release (Figure 6-2d). Condensation LH release is increased at temperatures above 0°C below the convective core due to the slightly higher LWC resulting from melting ice particles (Figure 6-1d). Increased heterogeneous ice formation significantly increases deposition LH release in the heterogeneous nucleation regime due to the growth of more numerous ice particles. The maximum positive (heating) value of
deposition LH is contained within the black contour line depicting strong updrafts (Figure 6-2e) indicating a connection between increased latent heat release and updraft intensity. Stronger downdrafts in the D1.2 case also correspond to greater negative (cooling) deposition LH from the increased sublimation of more numerous but smaller ice particles (Chapter 4 & 5; Gibbons et al., 2017).

Figure 6-1: Zonally averaged longitude slice plot of similar DCC structures within the Clean (row 1) and D1.2 (row 2) cases. Shaded colors: Liquid water content (LWC, column 1), Ice water content (IWC, column 2), and Vertical motion (column 3); Line contours, columns 1 & 2: vertical motion (solid black >1m/s; dashed blue < -0.1m/s); cloudiness threshold (dashed grey(black) line, >1e-6 kg/kg).
Figure 6-2: Zonally averaged longitude slice plot of similar DCC structures within the Clean (row 1) and D1.2 (row 2) cases. Shaded colors: Condensation/Evaporation Latent Heat (K/hr, column 1), Deposition/Sublimation Latent Heat (K/hr, column 2), and Freezing/Melting (1.e-3 K/hr, column 3); Line contours, columns 1 & 2: vertical motion (solid black >1m/s; dashed blue <-0.1m/s); cloudiness threshold (dashed grey, >1e-6 kg/kg).

Enhanced latent heat release in the dust cases results in stronger convective condensate outflow into the adjoining stratiform regime where they may grow to larger sizes, contributing to the increased upper level stratiform rain rates (Chapter 5; Min et al. 2009; Li and Min, 2010).
Figure 6-3 describes the changes in the dust and clean cases to horizontal and vertical mass flux related to in-storm circulations. Horizontal mass flux is defined as total water content multiplied by convergence/divergence. Vertical mass flux is defined as total water content multiplied by vertical motion. Increased heterogeneous ice formation and growth in the dust cases results in stronger vertical motion as indicated by a larger area of strong updraft velocities (> 1m/s). This is accompanied by stronger convergence and divergence within the convective core. The stratiform regime likewise has a correspondingly stronger influx of mass from the core. Divergence is stronger in the stratiform regime between 4km and 7km where downdraft intensity is also increased. The increased circulation of mass between the stratiform and convective core in the dust cases may be the source of the increased number of large liquid drops in the heterogeneous nucleation regime described in chapter 4 (Figure 4-8) and Chapter 5 (Figure 5-5).
Figure 6-3: Zonally averaged longitude slice plot of similar DCC structures within the Clean (column 1) and D1.2 (column 2) cases. Shaded colors: Horizontal condensate mass flux ($1.e-4$ g m$^{-3}$ s$^{-1}$, row 1), Vertical condensate mass flux (g m$^{-2}$ s$^{-1}$, row 2); Line contours, columns 1 & 2: vertical motion (solid black $>$1m/s; dashed blue $<$-0.1m/s); cloudiness threshold (dashed grey, $>$1e-6 kg/kg).
6.2.2 Convective and Stratiform Latent Heat Processes

The strong dynamic intensity of the convective core results in significant latent heat exchange within the updrafts and downdrafts as hydrometeors grow by diffusion and evaporate, respectively. Figure 6-4 describes the time evolution of different latent heat processes averaged over convective data points. Row 1 and row 2 describe latent heat released by liquid condensation and ice deposition processes, respectively. Row 3 describes the exchange of latent heat by freezing and melting processes. Row 4 describes the combined latent cooling resulting from evaporation and sublimation. Increased heterogeneous ice formation and growth in the dust cases reduces LWC at temperatures below 0°C, which decreases condensation heating in the heterogeneous nucleation regime. Conversely, the formation of more numerous ice particles in the dust cases significantly increases depositional heating, which in turn invigorates convective updrafts (Figure 6-7). The conversion of liquid drops to ice in the heterogeneous nucleation regime results in progressively reduced homogeneous drop freezing and latent heat release (Figure 6-4, row 3). Latent cooling from the melting process is reduced in the convective regime due to the formation of smaller particles which evaporate in the heterogeneous nucleation regime before melting can occur (Figure 6-4, row 4).
Figure 6-4: Time series of convective regime averaged vertical profiles of: Condensation latent heating (row 1), Deposition latent heating (row 2), Freezing/melting latent heating/cooling (row 3), and Evaporation/sublimation latent cooling (row 4).

Figure 6-5 describes time averaged plots of the latent heat processes (previously described in Figure 6-4) for both the convective and stratiform regimes. Changes to condensation and deposition heating in both regimes are similar, with smaller magnitudes in the stratiform regime. Overall, condensation heating decreases and deposition heating increases in the heterogeneous nucleation regime as IN concentration is increased in the dust cases. Likewise, evaporation/sublimation cooling is increased in both regimes. The greater mass of particles
transported into the stratiform regime increases particle melting and cooling compared to the Clean case. Latent cooling resulting from particle melting differs between the convective and stratiform regimes as IN concentration increases. In the convective core available water vapor is partitioned over a greater number of cloud ice particles which shifts the PSD to smaller sizes. Smaller particles are more likely to evaporate before melting which reduces the total cooling by melting below the core. Conversely, in the stratiform regime particles have grown to larger sizes by diffusional and/or collection processes resulting in greater cooling when the particles melt. The larger particles forming in the D1.2 case results in greater melting latent cooling compared to the D12 case despite the latter case’s greater total IWC.
Figure 6-5: Time averaged vertical profiles of convective (row 1) and stratiform (row 2) regime averaged: Condensation latent heating (column 1), Deposition latent heating (column 2), Freezing/melting latent heating/cooling (column 3), and Evaporation/sublimation latent cooling (column 4). Colors represent: Clean (black), D.12 (blue), D1.2 (red), D12 (green) cases, respectively.

Latent cooling by evaporation is affected by changes to both total hydrometeor mass and a shift to smaller particle sizes overall in the dust cases. Figure 6-6 describes the time averaged vertical profiles of evaporation for three different conditions within the convective and stratiform cloud regimes. Rain evaporation in the first column describes the evaporation of precipitation after falling outside the cloud body. Cloud edge evaporation in the second column describes
evaporation anywhere the cloud borders a non-cloudy pixel. Downdraft evaporation in the third column describes the evaporation occurring within data points featuring negative vertical motion. Convective rain evaporation is increased below the 0°C temperature level in the D.12 and D1.2 cases due to increased precipitation formation in the heterogeneous nucleation regime. The D12 case rain evaporation is reduced compared to the Clean case due to the overall smaller particle sizes which shifts precipitation formation to colder temperatures in the heterogeneous nucleation regime. Stronger vertical motion in the dust cases increases both the cloud edge and downdraft evaporation due to greater cloud area and more numerous small particles, respectively (Chapter 4; Gibbons et al., 2017). These changes are also seen in the stratiform regime at similar altitudes in the homogeneous nucleation regime. Stratiform rain evaporation is increased due to the increased formation of stratiform precipitation (mostly snow) in the dust cases (Chapter 5, Figures 5-8 and 5-9).
Figure 6-6: Time averaged vertical profiles of convective (row 1) and stratiform (row 2) regime averaged: Rain evaporation/sublimation latent cooling (column 1), Cloud edge evaporation latent cooling (column 2), and downdraft evaporation/sublimation latent cooling (column 3). Colors represent: Clean (black), D.12 (blue), D1.2 (red), D12 (green) cases, respectively.

Changes in latent heat profiles affect both vertical motion within the storm as well as the resulting convergence and divergence between the convective core and adjoining stratiform regime. Figure 6-7 describes the time averaged vertical profiles of total latent heat, wind convergence/divergence, updraft velocity, and downdraft velocity of the convective (row one) and stratiform (row two) regimes, respectively. Convective latent heat increases in the dust cases between 3km and 9km due to increased heterogeneous ice formation and subsequent growth.
This results in a more positive convergence value around 5km in the convective core and increased (more negative) divergence above 7km. Updraft velocity within the convective core increases in the dust cases due to greater latent heat release during ice particle diffusional growth. Compensatory downdrafts are also increased respectively. Stratiform latent heat is more negative overall in the dust cases due to increased evaporation of smaller particles forming locally and/or being transported from the convective core. Convergence within the stratiform regime is increased between 6km and 12km due to increased outflow from the convective core. Increased convergence also results in stronger stratiform downdrafts in the heterogeneous nucleation regime. This corresponds with increased stratiform divergence and increased convective convergence between 3km and 6km indicating and increased feedback between the two regimes.
Figure 6-7: Time averaged vertical profiles of convective (row 1) and stratiform (row 2) regime averaged: total latent heat (K h$^{-1}$, column 1), wind convergence (1.e-4 s$^{-1}$, column 2), updraft velocity (m s$^{-1}$, column 3), and downdraft velocity (m s$^{-1}$, column 4). Colors represent: Clean (black), D.12 (blue), D1.2 (red), D12 (green) cases, respectively.

6.2.3 Dynamical and Thermodynamical Regimes

Changes to microphysical processes are dependent on the underlying cloud dynamical and thermodynamical environment to allow for particle growth by diffusion as well as removal by sedimentation or evaporation. Figure 6-8 (row 1) describes the combined ice and liquid cloud water content (CWC) averaged with respect to temperature and vertical motion. Figure 6-8 (row
2) describes the combined rain, snow, and graupel water content (RWC) to represent the overall changes to precipitation formation. Increased heterogeneous ice formation and growth in the D.12 and D1.2 cases compared to the Clean case reduces CWC and increases RWC content at temperatures below 0°C. This is especially significant in the D1.2 case at temperatures below -40°C due to significantly reduced homogeneous freezing and IN depletion which results in a greater number of cloud ice particles being collected into large snow particles (Chapter 4 & 5, Gibbons et al., 2017). In the D12 case, CWC is most significantly increased between 0°C and -10°C due to significant heterogeneous ice formation and growth. CWC is also increased at temperatures below -40°C compared to the Clean case despite little homogeneous ice formation occurring (Chapter 4; Gibbons et al., 2017) due to increased vertical transport of cloud particles in the stronger updrafts (Figure 6-7). The formation of more numerous small ice particles in the D12 case delays the formation of precipitation to colder temperatures in the heterogeneous nucleation regime due to reduced collection efficiency (Chapter 5, Gibbons et al., 2017). As a result updraft RWC values are reduced between 0°C and -10°C compared to the Clean and other dust cases.

Figure 6-9 describes condensation (row 1), deposition (row 2), and freezing/melting (F-M, row 3) latent heat exchange averaged with respect to temperature and vertical motion as in Figure 6-8. Greater heterogeneous ice formation results in increased latent heat exchange by deposition/sublimation processes in the dust cases due to more numerous but smaller ice particles. Conversion of liquid water content to ice in the heterogeneous nucleation regime results in reduced latent heat exchange by condensation/evaporation processes as IN concentration is increased. Likewise, as fewer liquid drops are carried to the -38°C threshold, latent heating due to drop freezing processes is reduced. Conversely, deposition latent heat
release is increased as IN concentration is increased due to the greater ice formation and subsequent growth. The peak value of deposition LH (Figure 6-9) in the D1.2 and D12 cases correspond to locations of increased CWC and reduced RWC (Figure 6-8) due to the shifted vertical distribution of precipitation in the dust cases which results in increased diffusional growth of cloud ice particles.

**Figure 6-8:** Ice + Liquid Cloud (row 1) and Rain (row 2) water content (g m$^{-3}$) averaged with respect to temperature and vertical motion for the Clean and dust cases.
Figure 6-9: Condensation/evaporation (row 1), Deposition/sublimation (row 2), and Freezing/melting (row 3) latent heat averaged with respect to temperature and vertical motion for the Clean and dust cases.

To compare their physics-based latent heat retrieval with CRM derived latent heat, Li et al. (2013) described the fractions of deposition or freezing/melting latent heat to total latent heat in a temperature (thermodynamical) and storm height (dynamical) phase space. Figure 6-10 describes the fraction of deposition (row 1) and freezing/melting latent heat to the total latent heat for the Clean and dust cases. The deposition fraction of the Clean case is similar to the deposition fraction described by Li et al. (2013). Increased ice content and reduced liquid content in the dust cases results in progressively higher deposition latent heat fractions as IN concentration is increased. The fraction of freezing/melting (F-M) latent heat is initially increased in the D.12 case as a result of slightly increased vertical motion, which increases the
transport of liquid drops to homogeneous freezing temperatures. In the D1.2 and D12 cases, homogeneous freezing is reduced decreasing the F-M fraction. At temperatures above 0°C, the F-M fraction is increased in the D.12 and D1.2 cases due to the increased formation of large snow and graupel particles in the heterogeneous nucleation regime. While the D12 case F-M fraction is also increased at temperatures above 0°C compared to the Clean case, it is lower than the D1.2 case. This is due to increased formation of small ice crystals in the D12 case which shifts the formation of large snow and graupel to colder temperatures in the heterogeneous nucleation regime. Smaller crystals also sediment out more slowly and sublimate more readily than larger particles which reduces the number of particles melting above 0°C.

**Figure 6-10:** Row 1: Fraction of deposition/sublimation latent heat to total latent heat averaged with respect to temperature and cloud top temperature (CTT). Row 2: as row 1, for freezing/melting latent heat.
6.2.4 Effects on Total Domain Averages

To estimate the overall effects of increased IN concentration on latent heat processes; Figure 6-11 describes the time series of total latent heat averaged over all cloudy points in the domain (row 1). Figure 6-11 row 2 and row 3 describe the domain averages where vertical motion is either positive or negative, respectively. When latent heat is averaged with respect to both updrafts and downdrafts, latent heat values in the heterogeneous nucleation regime are decreased as IN concentration is increased from the D.12 case value. However, it is clear from prior figures that both latent heat exchange and vertical motion are increased in the dust cases. Domain averages of updraft and downdraft latent heat in the dust cases reinforce that both latent heating and cooling are increased in the heterogeneous nucleation regime as IN concentrations are increased in the dust cases.

**Figure 6-11:** Time series of whole domain (cloudy grid point) averaged vertical profiles of: total latent heat (row 1), updraft latent heat (row 2), and downdraft latent heat (row 3).
Figure 6-12 describes domain averaged time series dust case minus Clean case differences of wind shear (1-7km, 6-12a), relative humidity (T < 0°C, 6-12b), Average CAPE (1-7km, 6-12c), and average cold pool temperature (1-5km, 6-12d). Between hour 10 and hour 24 when convection is most intense, wind shear increases in the dust cases for greater IN concentrations as a result of stronger vertical motion and stronger convergence/divergence. Likewise, relative humidity at temperatures below 0°C is increased in the dust case due to increased evaporation/sublimation of the more numerous but smaller ice particles in the heterogeneous nucleation regime. Increased ice formation in the dust cases also affects average CAPE values and cold pool temperature although changes are not linear with increasing IN concentration.
**Figure 6-12:** Whole domain (cloudy grid point) averaged dust case minus Clean case time series of: (a) Wind shear (1 – 7 km), (b) Relative Humidity (T < 0°C), CAPE (1-5km), and Cold pool temperature (1-5km).

### 6.3 Conclusions

The MCS occurring on 08 March 2004 in the tropical eastern Atlantic, first described in Min et al. (2009) was simulated using the WRF model with a spectral-bin microphysical scheme. Ice nucleation parameters within the SBM were updated to connect heterogeneous and
homogeneous ice formation with a prognostic IN variable to investigate the effects of dust acting as IN. In Chapter 6, we focus on changes to latent heat processes and the subsequent impacts on atmospheric circulations and relative humidity.

Increased dust case heterogeneous ice formation and growth results in significant depositional latent heat release in the convective regime (Figure 6-4, Figure 6-5). This in turn invigorates convective updraft intensity and increases convergence/divergence in both the convective and stratiform regimes (Figure 6-3; Figure 6-7) and increases average wind shear over the 4th domain (Figure 6-12a). This results in increased condensate mass flux both vertically and horizontally into the adjoining stratiform regime (Figure 6-3b). The shift to smaller ice and graupel particle sizes in the dust cases results in increased sublimation cooling at cloud boundaries and in downdrafts (Figure 6-6), which increases average relative humidity at temperatures below 0°C (Figure 6-12b).

Increased conversion of liquid drops to ice particles in the heterogeneous nucleation regime, due to riming and/or drop freezing, results in the reduced contribution of condensation heating to total latent heating. Likewise, reduced homogeneous ice formation in the dust cases reduces latent heat release by drop freezing at the -38°C freezing level. However the increased ice water content in the HNR in the stratiform regime results in greater latent cooling near 0°C as more frozen particles melt (Figure 6-7; Figure 6-9; Figure 6-10). This is particularly evident in the D1.2 case due to the significant growth of large snow particles compared to the other dust cases after available IN have been depleted (Figure 6-8; Chapter 4: Figure 4-5; Figure 4-9).
Chapter 7

Summary and Future Work

7.1 Summary

The primary purpose of this thesis is to further the understanding of the impacts of ice-forming aerosols on deep convective clouds. Deep convective clouds have significant impacts on cloudiness, radiative transfer, precipitation and local and regional atmospheric circulations. Within this thesis, I have made improvements to the representation of heterogeneous ice formation in connection with prognostic IN concentrations within the WRF-SBM model. I have implemented an adjustable temperature and moisture content into the WRF-SBM’s simulated dust layer, in addition to differing number concentrations of IN and/or CCN. In addition, I have improved simulation of radar reflectivity from the model’s bin particle size distributions.

Specifically, the implementation of multiple heterogeneous ice formation mechanisms in connection with a prognostic IN variable is unique within our version of the WRF-SBM. It allows for ice formation to occur over the full range of temperatures within the heterogeneous nucleation regime. In addition, each mechanism can be activated and deactivated separately in order to test the sensitivity of the DCC’s microphysical processes to ice formed by each mechanism. An abundance of large frozen drops will affect the subsequent cloud and precipitation formation processes significantly differently than an equal number of small ice crystals. The current model setup produces results that correspond well with observed
microphysical and macrophysical changes resulting from the ingestion of dust into tropical DCCs.

The representation of the SAL in this study was also improved by allowing for adjustable temperature and moisture content within the simulated dust layer. While a dry air layer was present in the reanalysis data used to initialize the model, moisture content was overestimated compared to observations. Reducing moisture content by 5% significantly increased the D1.2 case’s agreement with observed reflectivity in both the convective and stratiform rain regimes. The adjustable dust layer opens further opportunities to test the sensitivity of cloud and precipitation formation processes for different added CCN and/or IN concentrations for a given dust layer temperature and moisture content.

7.2 Future Work

7.2.1 Improving the Representation of Ice Phase Processes in DCC

Aerosol related changes to initial ice formation have a significant impact on cloud and precipitation formation processes in DCC. Directly connecting ice formation with aerosol number concentration greatly improves our understanding of the microphysical and macrophysical changes occurring within the observed case study MCS (Min et al., 2009; Min and Li, 2010; Li and Min, 2010, Min et al., 2014). However, changes to the temperature range of primary ice formation also impacts the shape of the resulting ice crystals, otherwise known as the ice crystal habit (Bailey & Hallet, 2009). Crystal habit can have substantial impacts on ice crystal diffusional growth (Shaw and Mason, 1955; Pruppacher and Klett, 1997), particle terminal velocity (Mitchell et al., 1990; Mitchell, 1996; Mitchell & Heymsfield, 2005; Heymsfield et al., 2007), particle collision efficiency (Bohm, 1992; Wang and Ji, 2000; Hashino & Tripoli,
Ice crystal habit not constant and will evolve as the crystal grows under different temperature and supersaturation regimes, further affecting subsequent cloud and precipitation formation processes. Therefore, accurately representing ice crystal habits within numerical models is crucial to improving our understanding of ice phase processes in the atmosphere.

Numerical models incorporating an evolving crystal habit have been attempted before (Hashino and Tripoli, 2007, 2008, 2011a, 2011b; Misumi et al., 2010, Sulia et al., 2011; Harrington et al., 2013; Chen and Tsai, 2016). However, incorporation of an evolving crystal habit into a 3D cloud resolving bin-microphysics model and including crystal habit relevant effects on aggregation and riming processes has never been attempted. Implementing evolving ice crystal habit into the WRF-SBM will significantly improve the realistic representation of cloud and precipitation formation and growth processes from the initial nucleation of the ice particle, through the growth of the particle by diffusional and collision-collection processes, to the eventual removal of the particle from the atmosphere by sedimentation or evaporation. The following chapter describes an initial consideration of the potential effects of incorporating an evolving ice crystal habit into the WRF-SBM.

7.2.2 Ice Crystal Habit and Aspect Ratio Evolution

Ice crystals in nature are frequently irregular in shape and rarely form perfectly symmetrical shapes (Bailey & Hallet, 2009) However, the complexity of modeling such complex shapes in a 3D cloud resolved model is currently unfeasible. Instead, prolate and oblate spheroids can be used to provide a first order approximation of the evolution of a crystals basal and prism face radii as an improvement over the use of equivalent density spheres (Sulia et al., 2011). The
ratio of the radii of a crystal’s basal face (c axis) to its prism face (a axis) is termed the crystal’s aspect ratio (φ=c/a). The use of the aspect ratio also allows for the estimation of secondary crystal habits which are affected by more complicated factors than temperature alone (Chen & Lamb, 1994). An ice crystal’s eventual habit is affected by both the temperature and supersaturation conditions experienced by the crystal during diffusional growth. The temperature dependent component is known as the inherent growth ratio (IGR) and has been derived from a combination of experimental and observational values (Chen & Lamb, 1994).

Figure 7-1a describes the appropriate values of the IGR for each model level to demonstrate the effects of the location of initial ice formation on the resulting primary crystal habit. Within the heterogeneous nucleation regime, the crystal habit transitions through plate, column, plate and column shapes as temperature decreases. Spherical ice particles occur at ~ -4°C, -9°C, and -21°C due to both crystal axes growing at similar rates (Figure 7-1b). The specific aspect ratio of a crystal will be affected by the initial size of the crystal when it is first formed. Figure 7-1b describes the a-axis (orange line) and c-axis (purple line) lengths of an ice particle with a spherical equivalent radius (R_{eq}) of 20um having grown from an initial radius (R_i) of 2um (solid line) or 10um (dotted line). The results were calculated based on Chen & Tsai (2016, eq. 8) using the IGR in described in Figure 1a. As noted by (Sheridan et al, 2009; Sulia et al., 2011; Chen & Tsai, 2016) the smaller initial crystal radii results in more extreme aspect ratios (Figure 7-1b; Figure 7-2a).
Figure 7-1: (a) Vertical Profile of inherent growth ratio based on Chen & Lamb (1994). (b) Vertical profile of a (orange) and c (purple) axis lengths for ice crystals with spherical equivalent radii (Req) of 20um. Initial Radii (Ri) of particles at formation assumed to be 2um (solid line) or 10um (dotted line).

More extreme aspect ratios will result in a greater 2D crystal area for a given particle mass which affects particle fall rates (Heymsfield & Kajikawa, 1987; Mitchell, 1996; Mitchell & Heymsfield, 2005), particle collision efficiencies (Hashino & Tripoli, 2011a,b; Jensen & Harrington, 2015), and polarimetric radar (Ryzhkov et al., 2011; Hogan et al., 2012). Figure 3 describes the aspect ratio and 2D particle area of an ice crystal at bin #22 (Req~250um) for different initial bin numbers (providing initial radii, Ri). Aspect ratio was calculated based on the equations used in Figure 7-1b, with the IGR for the column and plate crystals set equal to 2.0 and 0.5, respectively. As noted in Figure 1b, aspect ratio is more extreme for smaller initial crystal sizes. Figure 7-2b calculates the equivalent 2D area (Jensen & Harrington, 2015) of the crystal,
assuming a density of bulk ice. Values for a spherical particle (purple line) are provided for reference. The resulting area of both plates and columns are greater than the equivalent spherical value, with plates obtaining the greatest potential area (+900%). Therefore, changes to the initial location (temperature range) and particle size of ice crystal formation can have a significant impact on ice microphysical processes which is poorly represented in current 3D cloud models that assume only spherical particles or fixed crystal habits.

**Figure 7-2:** (a) Aspect ratio of ice crystals at bin # 22 (Req ~250) for different initial bin numbers (Ri). (b) Equivalent 2D particle area for the calculated aspect ratio. Particle density assumed to be that of bulk ice. Colors indicate Column (black), Plate (red), or spherical (purple) ice crystals, respectively.

Greater concentrations of IN in dust case simulations will result in increased ice formation in the heterogeneous nucleation regime while also reducing ice formation by
homogeneous drops freezing. This affects the initial temperature regime in which ice crystals begin their initial growth and, correspondingly, the initial shape of the ice crystal. The specific ice formation mechanism will also have an effect on the eventual crystal habit due to the different initial radii of ice particles formed by drop freezing as opposed to deposition nucleation. While drop freezing may form ice over the range of the liquid bin PSD, deposition nucleation in the WRF-SBM always forms ice crystals with a radii of 2um (the smallest bin).

Figure 7-3 describes the vertical profile of total ice formation number (cm\(^{-3}\) Log\(_{10}\)) summed over the entire 33 hour simulation time. Row 1 and row 2 describe convective and stratiform regime ice formation, respectively. Column 1 and 2 describe the ice formation resulting from all drop freezing mechanisms (Homogeneous, Immersion, and Contact), and deposition nucleation, respectively. As IN concentrations are increased in the dust cases, drop freezing is reduced due to the increased conversion of liquid drops to ice by riming processes, immersion/contact drop freezing, and/or Bergeron evaporation (Chapters 4 & 5). Deposition nucleation is increased over the vertical range of the heterogeneous nucleation regime which results in the significant increase in the number of small ice crystals as noted in Chapter 4 and 5. These small crystals will experience more rapid growth compared to initially larger particles (Sheridan et al., 2009; Sulia et al., 2011) which may affect water vapor availability within the cloud and potentially change the partition between liquid and ice due to Bergeron evaporation.
Figure 7-3: Vertical profile of ice formation number resulting from drop freezing (Homogeneous, Immersion, and Contact; Column 1) and deposition nucleation (Column 2) mechanisms summed over entire 33 hour simulation. Row 1 and 2 describe convective and stratiform ice formation, respectively. Colors represent: Clean (black), D.12 (blue), D1.2 (red), D12 (green) cases, respectively.
7.2.3 Aggregate Density and Crystal Habit

Observations suggest that mature and or rimed aggregate particles tend to be horizontally aligned oblate spheroids with an overall aspect ratio varying between 0.8 and 0.6 (Hashino & Tripoli, 2011a,b; Hogan et al., 2012). The mass-dimensional relationship for aggregate particles is dependent on the crystal habit of the constituent crystals and the degree of riming (Mitchell et al., 1990; Hashino & Tripoli, 2011b; Kuo et al., 2016). This affects the density of an aggregate particle of a given mass, which in turn affects particle fall rates (Barthazy & Schefold, 2006) and radar reflectivity (Ryzhkov et al. 2011; Kneifel et al., 2015; Leinonen & Moisseev, 2015). Figure 7-4 describes the calculated particle radius and density of aggregate particles for a given particle mass (bin number) assuming different habits for the component crystals and/or significant riming. Calculations are based on the mass-dimension relationships described in Mitchell et al. (1990) for the appropriate crystal type. In general, aspect ratios near 1.0 and/or the presence of significant riming will result in a smaller and thereby denser aggregate particle for a given mass. This is due to closer packing of crystals (in the former case) or rimed mass filling in air space within the aggregate (in the latter case). The difference between aggregates composed purely of plates and those composed of a mixture of different crystal habits is fairly subtle. In this case the effect is similar to the presence of riming mass within the aggregate as crystals with less extreme habits may fill in gaps within the matrix formed by crystals with more extreme aspect ratios.
Figure 7-4: Calculated aggregate radii (a) and density (b) for given particle mass (bin number) accounting for constituent crystal habit and/or particle riming. Colors represent: collections of long columns (LC, purple); collections of rimed large columns (LCR, blue); collections of short columns (SC, green); Aggregates of plates (AggrP, yellow); Aggregates of mixed crystal types (AggrM, orange); heavily rimed aggregates (AggrR, red).

Increased heterogeneous ice formation such as those simulated in our dust cases will affect both the rates of aggregate formation and the frequency of riming. Figure 5 describes the convective riming rates for snow (a) and graupel (b) particles for the Clean and dust cases. The change of ice particle number during riming (Aggregate Number) and the dust case minus Clean case difference of aggregate number are described in Figure 7-5c and 7-5d, respectively. The increased heterogeneous ice formation in the dust cases results in more frequent riming and increased graupel formation at temperatures near 0°C. At colder temperatures, riming rates are reduced due to fewer drops remaining unfrozen at these temperatures. For snow particles to gain
mass by riming without being converted into graupel, the liquid water content must be lower than the snow water content. Therefore snow particle riming rates in Figure 7-5 are lower than those of graupel particles despite the larger inherent capture radius of a snow particle compared to the denser graupel particles.

**Figure 7-5:** Time averaged vertical profiles for convective regime: Snow riming rate (a); Graupel riming rate (b); Aggregate number (c); Dust case minus Clean case difference of aggregate number (d). Colors represent: Clean (black), D.1 (blue), D.1.2 (red), D12 (green) cases, respectively.
7.2.4 Partial melting and Radar Reflectivity

Large snow particles have a significant effect on radar reflectivity due to their large radii for a given particle mass. Partial melting of these snow particles are frequently represented by a radar "bright band" in observations. The shape and density of snow are significantly affected by partial melting compared to frozen particles that are both denser and typically more spherical in shape, such as graupel (Fabry & Zawadzki, 1995; Fabry & Szyrmer, 1999; Szyrmer & Zawadzki, 1999; Oraltay & Hallett, 2005; Zawadzki et al., 2005). Figure 7-6 describes calculated radar reflectivity for snow particles with masses ranging from bin #18 to bin #33 for different melted fractions (0% to 60%). For simplicity sake, snow density is set to 0.1 (g m\(^{-3}\)) for all bins unless otherwise noted and particles are assumed to be spherical. To isolate the effects of changes to particle density and dielectric factor on reflectivity, Figure 6a uses the appropriate melted radii for all melted fractions listed. Density and dielectric constant (Chapter 3; Ryzhkov et al., 2011) are adjusted to reflect the appropriate melted fraction. Figure 7-6b describes reflectivity calculated for fixed density and dielectric constants and using density adjusted particle radii. Partial melting of snow particles affects radar reflectivity both positively (density and dielectric constant) and negatively (reduced radii) which complicates accurately reproducing the radar bright band. For a non-spherical particle, partial melting can also result in changes to the overall particle shape which will further affect reflectivity during melting. While previous studies using bin microphysics have tracked melted fraction of ice particles (Khain et al., 2004, for example), accounting for partial melting and evolving ice crystal habit has the potential to significantly improve understanding of the radar bright band.
Figure 7-6: Radar reflectivity for partially melted aggregates. (a) Values calculated using changes to particle density and dielectric constant. Radii used is equivalent melted radii. (b) Values calculated using fixed particle density (0.1 g cm\(^{-3}\)) and radii adjusted for partial melting.
Chapter 8

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