The Microphysical Structure of Mesoscale Convective Systems

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Abstract

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Mesoscale convective systems (MCSs) are large, long-duration complexes of clouds that are composed of a mixture of convective and stratiform components united by a mesoscale circulation. By developing an innovative spatial compositing technique that combines dual-polarimetric and Doppler radar data obtained during the Dynamics of the Madden-Julian Oscillation/ARM MJO Investigation Experiment (DYNAMO/AMIE), it is shown that hydrometeors are systematically organized around the mesoscale airflow patterns in MCSs in manner that is consistent with their known dynamical structure. Nine different hydrometeor types are identified by applying the National Center for Atmospheric Research (NCAR) particle identification (PID) algorithm to dual-polarimetric data obtained from the NCAR S-PolKa radar. The organization of these hydrometeors relative to airflow through MCSs is determined by
simultaneously examining Doppler-radar-observed air motions and PID data. In convective cores, moderate rain occurs within the updraft core, where the rapidly rising air prevents hydrometeors from growing significantly. The heaviest rain and narrow, shallow regions of graupel/rimed aggregates are located just downstream of the updraft core, where the convective downdraft is likely located. Wet aggregates are located slightly further downstream from the updraft core in a narrow layer just below the 0°C layer, where the vertical velocities are likely less intense. The upper-levels of the convective core, where there is a lot of turbulence, are dominated by dry aggregates. Small ice crystals are located along the cloud edges. Within the stratiform region the rain intensity systematically decreases with distance from the center of the storm. Descending from cloud top small ice crystals, dry aggregates, and wet aggregates are sequentially layered in a manner consistent with the gradual gravitational setting observed in the upper portions of the stratiform region. Additionally, pockets of graupel/rimed aggregates are occasionally observed just above the wet aggregate layer. It is suggested that these graupel/rimed aggregates could result from localized wind-shear-induced turbulence, previous convective cells, and/or small, embedded convective cells. While previous studies have found evidence of these spatial hydrometeor patterns, this dissertation is the first to analyze Doppler-radar-observe air motions simultaneously with the PID data and show that these are patterns are systematically organized with respect to the mesoscale circulation of MCSs. Thus, this work builds upon a 50 year tradition of using the latest radar technology to advance our understanding of the fundamental nature of tropical oceanic MCS.

While the PID is traditionally interpreted as an indication of the dominant hydrometeor type within a volume of air sampled by a radar, this dissertation takes advantage of the fact that the frozen hydrometeors identified by the PID methodology can be interpreted in terms of the
microphysical processes producing the ice particles. Using this microphysical interpretation of the PID and constraining simulations to have the same mesoscale circulations as observations, the second part of this dissertation investigates whether numerical simulations can replicate the microphysical patterns observed in the S-PolKa data in a manner that is consistent with previous theoretical and laboratory studies. These simulations used three routinely available microphysical parameterizations. The simulated mesoscale airflow patterns were free to interact with the model microphysics, but the air motions were constrained to observations via assimilation of the S-PolKa radial velocity data. Broadly speaking, the simulated ice microphysical patterns were consistent with each other, with radar observations, and with previous theories and laboratory studies. All suggest that the ice microphysical processes in the midlevel inflow region are characterized by deposition anywhere above the 0°C level where upward vertical velocity is present, aggregation at and above the 0°C level, riming near the 0°C level, and melting at and below the 0°C level. Despite these broad similarities, the simulated ice microphysical patterns substantially differed in details from observations and previous theoretical and laboratory studies. Each simulation was characterized by riming and aggregation occurring over too deep of a layer and riming occurred too frequently. Additionally, details of the simulated ice microphysical patterns always differed among the parameterizations; no two parameterizations consistently produced similar details in every ice process considered. These discrepancies likely factored into creating substantial reflectivity differences among the parametrizations and with observations, which suggests that reliable consistent simulations will not be achieved until the parameterized representation of microphysical processes is improved. As a whole, this dissertation advances our understanding of the fundamental nature of tropical oceanic MCSs and provides important insights into the relationship between the
dynamical and microphysical patterns within these storms from an observational and modeling perspective.
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GLOSSARY

AMIE: ARM MJO Investigation Experiment

ARM: Atmospheric Radiation Measurement Program of the U.S. Department of Energy

WRF-ARW: Advanced Research version of the Weather Research and Forecasting model

CNES: Centre National d’Etudes Spatiales

DA: Dry aggregates

DYNAMO: Dynamics of the Madden-Julian Oscillation field experiment

EnKF: Ensemble Kalman filter

G/R: Graupel/rain

G/RA: Graupel/rimed aggregates

GARP: Global Atmospheric Research Programme

GATE: GARP Atlantic Tropical Experiment

GMT: Greenwich Mean Time

H/R: Hail/rain

H: Hail

HI: Horizontally-oriented ice crystals

HR: Heavy rain

KDP: Specific differential phase

LDR: Linear depolarization ratio

LR: Light rain
MISMO: Mirai Indian Ocean Cruise for the Study of the MJO-Convection Onset

MJO: Madden-Julian Oscillation

MOR: Morrison 2-moment microphysics parameterization

MR: Moderate rain

MY: Milbranbt-Yau double-moment microphysics parameterization

NCAR: National Center for Atmospheric Research

NCAR EOL: Earth Observing Laboratory of NCAR

PID: Particle identification algorithm

RHI: Range-height indicator scan; displays radar data as a function of radial distance from radar and height

SI: Small ice particles

S-PolKa: Dual-polarimetric radar used during DYNAMO / AMIE, owned and operated by NCAR EOL.

TOGA COARE: Tropical Ocean – Global Atmospheric Coupled Ocean Atmosphere Response Experiment

UTC: Coordinated Universal Time; also known as GMT.

WA: Wet aggregates

WD: WRF double-moment 6-class microphysics parameterization

WRF: Weather Research and Forecasting model

Z_{DR}: Differential reflectivity

\rho_{HV}: Correlation coefficient
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DEDICATION

To my parents, who provide endless support and love
CHAPTER 1

DISSERTATION INTRODUCTION

Tropical mesoscale convective systems (MCSs) and their relationship with the large-scale atmospheric circulation, have constituted an active area of research for nearly 50 years. Mesoscale convective systems are an ensemble of convective and stratiform clouds that interact synergistically to create a complex that develops circulations that are larger than any of its individual components. These cloud systems are a crucially important feature of the tropical cloud population and global circulation. MCSs account for 40-60% of the precipitation over the tropical oceans (Houze, 2014) and deep convection accounts for nearly 50% of the upper-level cloud cover in the tropics (Lou and Rossow, 2004; Mace et al., 2006). They significantly impact the global circulation through their radiative fluxes, momentum fluxes, and latent heat (e.g. Schumacher et al., 2004; Mechem et al. 2006). This dissertation represents an important advancement in our understanding of MCSs and their relationship to the large-scale circulation by demonstrating how hydrometeors and ice microphysical processes are organized within these storms.

Photographic analysis conducted by Malkus and Riehl (1964) and early satellite imagery (e.g. Anderson et al., 1966) provided the first evidence that MCSs are a central feature of the tropical oceanic cloud population. However, in the 1960s and early 1970s, it was unclear why these large cloud systems existed (Frank, 1970). At that time, it was assumed that MCSs interacted with the
large-scale circulation through narrow “hot towers.” Riehl and Malkus (1958) proposed that deep convective cores in the deep tropics maintain the global circulation by transporting mass and energy directly from the boundary layer to the upper troposphere. The stratiform region of MCSs was assumed to be a dynamically inactive cirrus cloud shield whose only role was to protect the “hot tower” in the convective core (see the historical discussion by Houze, 2003). However, this “hot tower” view of MCSs was beginning to be challenged. Rawinsonde and aircraft data collected 1000 miles south of Hawaii during the Line Islands Experiment in 1967 indicated that these cloud systems had mesoscale circulations larger than any individual convective or stratiform entity within the MCSs (Zipser, 1969).

The Global Atmospheric Research Programme (GARP) Atlantic Tropical Experiment (GATE) revolutionized the understanding of MCSs and their role in the global circulation. GATE occurred in the tropical eastern Atlantic Ocean in 1974 and is the largest atmospheric science field campaign in history with 72 countries, 40 ships, and 12 aircraft participating (Kuettner, 1974). Four of the ships had radars onboard, which provided the first quantitative radar reflectivity data of tropical oceanic MCSs. These radars did not have Doppler or dual-polarization capability, but they were state-of-the-art radars at that time. By measuring the full three-dimensional radar reflectivity structure, these radars provided insight into the horizontal and vertical structure of precipitation, including MCSs. Using this radar data, Houze (1977) and Houze and Cheng (1977) showed that the stratiform component of MCSs over the tropical ocean accounted for ~40% of the precipitation from the MCSs. Gamache and Houze (1982) and Houze and Rappaport (1984) used GATE sounding and aircraft data to show that the stratiform region had upward motion aloft and downward motion in the lower troposphere. Using these results as well as results from the 1977-78 Monsoon Experiment (MONEX, described by Johnson and Houze, 1987), Houze (1982)
showed that the vertical motion profile and rainfall from tropical oceanic MCSs combine with radiative heating anomalies in the widespread stratiform cloud shield to create a top-heavy heating profile characterized by maximal heating at upper-levels and weak cooling at low-levels.

The potential significance of the top-heavy heating profile as seen in GATE and MONEX for the global circulation was demonstrated by Hartmann et al. (1984) and was part of the motivation for the design of the Tropical Rainfall Measurement Mission (TRMM) satellite (Simpson et al., 1988). Hartmann et al. (1984) obtained a more realistic vertical structure of the mean tropical circulation when equatorial convection was assumed to have a top-heaving heating profile. The conclusion that the top-heavy heating profile in the tropics is important to the global circulation by Hartmann et al. (1984) and other studies made it clear that an improved understanding of the structure and variability of latent heating in the tropics and its impact on the global circulation was necessary. The tool designed to address this necessity was the Tropical Rainfall Measurement Mission (TRMM, Simpson et al., 1988) satellite. Operating in a low-earth orbit, the TRMM satellite provided data about the distribution, variability, and vertical structure of tropical and subtropical precipitation from 1998 through 2015. The three-dimensional quantitative radar data obtained by the TRMM Precipitation Radar over the entire tropics has allowed the proportions of convective and stratiform precipitation to be determined throughout the tropical latitudes (Schumacher et al., 2003). Using data from the Precipitation Radar aboard TRMM, Schumacher et al. (2004) confirmed the importance of the stratiform component of MCSs for the global circulation and for features such as ENSO. More recently, Barnes and Houze (2013) and Barnes et al. (2015) used the TRMM data to show that the stratiform fraction of the convective population and its associated heating profile varies with the phase of the Madden-Julian Oscillation (MJO).
Tropical Ocean-Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE) in the West Pacific Ocean region was another milestone in understanding tropical oceanic MCSs because it applied Doppler radar technology to tropical oceanic MCSs for the first time. In GATE and MONEX air motions could only be inferred from soundings and aircraft flight-track wind data. However, during TOGA COARE, both airborne and shipborne Doppler radars were used to investigate the air motions within MCSs. From the airborne Doppler radar data, Mapes and Houze (1995) determined the vertical profiles of divergence in both the convective and stratiform regions of MCSs and, using shallow water equations, showed how the effects of the convective and stratiform heating profiles propagate differently through the large-scale environment. Kingsmill and Houze (1999a) also used the airborne Doppler radar data obtained in TOGA COARE. However, they used the data to demonstrate how air flows through MCSs in distinct three-dimensional layers. They found that the convective cores in MCSs are characterized by a layer of air steeply rising out of the boundary layer and diverging at cloud top. Airflow within the stratiform region is dominated by the midlevel inflow, which is a layer of air that converges near the bottom of the anvil and gradually descends towards the center of the storm. These airflow patterns are consistent with those derived by Zipser (1969) and simulated by Moncrieff (1992). However, the airflow simulated by Moncrieff (1992) was two-dimensional and Kingsmill and Houze (1999a) found that the airflow in MCSs is highly three-dimensional. Houze et al. (2000) obtained further insight from the TOGA COARE shipborne radars on how the mesoscale air motions occurred in different parts of the MJO circulation pattern and hypothesized that MCSs affected the momentum budget differently in different parts of the MJO circulation pattern. These hypotheses were validated in a modeling study by Mechem et al. (2006), who
demonstrated that the mesoscale airflow associated with MCSs impacts the large-scale momentum budget.

The recent Dynamics of the Madden-Julian Oscillation /ARM MJO Initiation Experiment (DYNAMO/AMIE), which took place in the equatorial Indian Ocean in the winter of 2011-2012, brought yet another level of radar technology to observations of tropical oceanic MCSs. While the underlying objective of DYNAMO/AMIE was to understand how the Madden-Julian Oscillation (MJO) initiates in the Indian Ocean, its diverse dataset can be used to gain insight into the tropical oceanic cloud population in general, of which MCSs are an especially important component (e.g. Barnes and Houze, 2013; Zuluaga and Houze, 2013; Barnes and Houze, 2015). An important feature of this project was that it was one of the first projects to apply dual-polarization radar technology to tropical oceanic MCSs. Dual-polarimetric radars emit and receive vertically and horizontally polarized pulses, which enables them to calculate additional moments of the particle size distribution that indicate the physical characteristics of particles within a volume of air sampled by the radar. These physical characteristics are indicative of the hydrometeors within the radar sample volume and the microphysical processes acting on them. Thus, dual-polarization technology used in DYNAMO/AMIE has allowed the microphysical characteristics of MCSs to be inferred. This microphysical data not only increases our understanding of MCSs, but it provides insight into the role of MCSs in the global circulation since these processes modify the radiative and latent heating structure of the atmosphere.

Knowledge of the organization of microphysical processes is important since these processes are linked to the precipitation, air motions, and heating profiles within convection. For example, buoyancy is modified as microphysical processes emit and absorb latent heat, which, in turn, contributes to the development and maintenance of vertical air motion (e.g. Szeto et al., 1988; Tao
et al., 1995; Adams-Selin et al., 2013). Additionally, the latent heat released and absorbed during microphysical processes activates teleconnections that alter the global circulation (e.g. Hartmann et al., 1984; Schumacher et al., 2004). Furthermore, microphysical processes impact the pattern of radiative heating within convection. Once stratiform precipitation is formed, ice microphysical processes modify radiative heating, which can increase instability, cause turbulence, and extend the lifetime of stratiform precipitation and its associated anvil cloud (e.g. Webster and Stephens, 1980; Chen and Cotton, 1988; Churchill and Houze, 1991; Tao et al., 1996).

Studies such as Chen and Cotton (1988) have demonstrated that skillful mesoscale simulations require accurate representations of microphysical processes, latent heating, and radiative transfer and their interactions. Thus, it is important that that the research community knows how microphysical processes are organized within observed convection and accurately represents these microphysical patterns in simulated convection. Studies including Leary and Houze (1979b), Houze (1981; 1989), Houze and Churchill (1987), and Braun and Houze (1994) either used conventional radar data, aircraft data, and/or deductive reasoning to develop conceptual models that comprehensively describe the spatial pattern of microphysical processes within MCSs. While these conceptual models suggest that microphysical processes are systematically associated with the kinematic structure of MCSs, the observational evidence of the link was weak. Additionally, knowledge of how three-dimensional, full-physics simulations spatially organize microphysical processes is limited. Caniaux et al. (1994) is one of the only studies that has shown where specific microphysical processes occur within simulated convection. However, their study used an idealized two-dimensional model that prevented dynamical and microphysical processes from interacting. Studies such as Donner et al. (2001) show the spatial structure of latent heat associated with specific microphysical processes, which provide some indication of the spatial organization
of processes that change the phase of water occur (e.g. deposition, riming, melting). However, these studies do not provide any information about the spatial organization of processes that do not change the phase of water (e.g. aggregation) in simulations. This dissertation resolves these shortcomings in our knowledge of the spatial pattern of microphysical processes from both observational and numerical simulation perspectives and suggests that the spatial pattern of microphysical processes within MCSs are linked to the kinematic structure of these storms.

Chapter 2 examines the precipitation (using reflectivity as in GATE) and the air motions (using radial velocity as in TOGA COARE) and combines them with microphysical processes in order to demonstrate that hydrometeors are systematically organized with respect to their classic convective updraft and midlevel inflow structures in MCSs. Thus, Chapter 2 provides direct observational evidence of how hydrometeors, and the microphysical processes acting on them, relate to the canonical kinematic structure that was first discussed by Zipser (1969) and elaborated by Moncrieff (1992) and Kingsmill and Houze (1999a).

Chapter 3 uses the microphysical patterns observed in Chapter 2, to evaluate three routinely available microphysical parameterizations for their ability to accurately represent the spatial pattern of microphysical processes relative to the midlevel inflow in MCSs. While the spatial pattern of microphysical processes is vital in order to accurately simulate convection at the local and global scales, little was known about how numerical models organize these processes within simulated convection prior to this dissertation. Thus, Chapter 3 contains important insights into how the next generation of microphysical parameterizations should be developed. As a unit, this dissertation directly builds upon a framework that has been developed over the last 50 years by demonstrating that hydrometeors and microphysical processes are systematically organized around the kinematic structure of these storms. This new insight into the vertical structure of
tropical oceanic MCSs provides a better understanding of the association between microphysical and dynamical processes in these storms.
CHAPTER 2

PRECIPITATION HYDROMETEOR TYPES RELATIVE TO THE MESOSCALE AIRFLOW IN MATURE OCEANIC DEEP CONVECTION OF THE MADDEN-JULIAN OSCILLATION

Composite analysis of mature near-equatorial oceanic mesoscale convective systems (MCSs) during the active stage of the Madden-Julian Oscillation (MJO) show where different hydrometeor types occur relative to convective updraft and stratiform midlevel inflow layers. The National Center for Atmospheric Research (NCAR) S-PolKa radar observed these MCSs during the Dynamics of the Madden-Julian Oscillation/Atmospheric Radiation Measurement-MJO Investigation Experiment (DYNAMO/AMIE). NCAR’s particle identification algorithm (PID) is applied to S-PolKa’s dual-polarimetric data to identify the dominant hydrometeor type in each radar sample volume. Combining S-PolKa’s Doppler-velocity data with the PID demonstrates that hydrometeors have a systematic relationship to the airflow within mature MCSs. In the convective region: moderate rain occurs within the updraft core; the heaviest rain occurs just downwind of the core; wet aggregates occur immediately below the melting layer; narrow zones containing graupel/rimed aggregates occur just downstream of the updraft core at midlevels; dry aggregates dominate above the melting level; and smaller ice particles occur along the edges of the convective zone. In the stratiform region: rain intensity decreases toward the anvil; melting aggregates occur in horizontally extensive but vertically thin regions at the melting layer; intermittent pockets of graupel/rimed aggregates occur atop the melting layer; dry aggregates and small ice particles occur sequentially above the melting level; horizontally-oriented ice crystals occur between –10°C to –
20°C in turbulent air above the descending midlevel inflow, suggesting enhanced depositional growth of dendrites. The organization of hydrometeors within the midlevel inflow layer is insensitive to the presence or absence of a leading convective line.

Publication Reference:
2.1 Introduction

Mesoscale convective systems (MCSs) are broadly defined as cloud systems whose contiguous precipitation spans at least ~100 km in one direction (Houze, 2004). These cloud systems are comprised of small, intensely precipitating convective regions and expansive stratiform regions that have a relatively steady but reduced precipitation rate. If the convective cells are organized into a line ahead of a moving stratiform region, the MCS is referred to as a stratiform region with a leading-convective line. The leading line is commonly called a squall line. If the convective cells are embedded within the MCS, the storm is said to contain a stratiform region without a leading-convective line. Leary and Houze (1979a) and Houze and Betts (1981) referred to these two types of MCSs as “squall clusters” and “non-squall clusters”, respectively. The convective and stratiform portions of MCSs are also characterized by distinct kinematic structures. Using an idealized, steady state two-dimensional numerical simulation with prescribed environmental instability and vertical wind shear, Moncrieff (1992) suggested that air moves through an MCS in coherent layers. Kingsmill and Houze (1999a) confirmed this layered airflow through a dual-Doppler analysis of airborne radar data obtained during the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE) in the west Pacific Ocean. Their results indicate that mature convective regions are characterized by a relatively deep surface convergent layer that steeply rises until it diverges near cloud top. Mature stratiform regions are distinguished by a midlevel inflow layer that converges beneath the anvil and gradually slopes downwards toward the center of the MCS. An upper-level mesoscale sloping updraft layer is located above the midlevel inflow layer.
These airflow patterns influence various aspects of the storm, including microphysical processes and the locations of different types of hydrometeors in three-dimensional space. More rapidly falling, denser hydrometeors reach the surface much closer to the convective updraft core. More slowly falling particles are advected farther downstream and help create the stratiform portion of the MCS. Aircraft probes are capable of determining hydrometeors and their microphysical properties, including their bulk water and ice content. In conjunction with airborne radars, these probes have been used to relate the microphysical and kinematic fields (e.g. Zrnić et al., 1993; Hogan et al., 2002; Bouniol et al., 2010). However, probe data has limited temporal and areal coverage since it is restricted to the aircraft’s flight path. Additionally, while airborne probes are capable of sampling frozen, mixed phase, and liquid hydrometeors, it is nearly impossible to simultaneously sample all three phases for a long period of time. Dual-polarimetric radars have the benefit of being able to identify microphysical and hydrometeor features continuously over broad three-dimensional spaces for long durations.

Previous radar studies have provided insight into the relationship between the kinematic and microphysical structure in midlatitude and tropical land regions. Evaristo et al. (2010) approximated the three-dimensional wind field of a West African squall line and compared it to the vertical structure of the hydrometeors identified by the PID. Höller et al. (1994) and Tessendorf et al. (2005) both used a PID as a tool to understand hail trajectories and growth processes in supercell thunderstorms. However, relatively little is known about these kinematic and microphysical relationships in tropical, oceanic regions. While the kinematic structure of mature MCSs is fundamentally similar throughout the globe (e.g. Zipser, 1977; Keenan and Carbone, 1992; LeMone et al., 1998), differences in the thermodynamic profile, aerosol content, and convective intensity (e.g. Zipser and LeMone, 1980; LeMone and Zipser, 1980) cause the
hydrometeor structure in tropical, oceanic MCSs to differ from midlatitude and terrestrial MCSs. Additionally, Houze (1989) demonstrated that the convective regions of mesoscale systems have different vertical velocity profiles in different tropical oceanic regions. Thus, the hydrometeor structure of MCSs likely differs between different oceanic regions. Relatively few studies have been conducted in the central Indian Ocean. It is important to resolve this knowledge gap and increase our understanding of precipitation processes in mature MCSs associated with the Madden-Julian Oscillation (MJO) in the central Indian Ocean, so that accurate parameterizations can be developed and the validity of numerical simulations can be more rigorously assessed.

The National Center for Atmospheric Research (NCAR) S-PolKa radar was deployed during the Dynamics of the Madden-Julian Oscillation/Atmospheric Radiation Measurement-MJO Investigation Experiment (DYNAMO/AMIE) in the equatorial Indian Ocean to document the structure and variability of the cloud population associated with the MJO (Yoneyama et al., 2013). The dual-polarimetric and Doppler capabilities of this radar enable this dissertation to directly investigate the association between the airflow and hydrometeors within mature tropical, oceanic MCSs.

The dual-polarimetric radar signatures of different hydrometeors are so complex that manual analysis is prohibitively time consuming for large samples of data. In order to aid in the analysis of this complicated data, NCAR has developed a particle identification algorithm (PID) that is applied to dual-polarimetric data to identify the most likely dominant hydrometeor from a given volume of radar data (Vivekanandan et al., 1999). Rowe and Houze (2014) composited PID data collected by the S-PolKa radar during DYNAMO/AMIE and investigated how the frequency and vertical profile of hydrometeors varied between three active periods of the MJO. They examined MCSs and smaller storms, which they called sub-MCSs. They concluded that the frequency of
different hydrometeor types vary with storm size and MJO active event. However, the mean shape of the vertical profile showed little variation, differing only according to whether they were located in the convective or stratiform portions of the precipitating cloud systems. Given that Rowe and Houze (2014) only considered the mesoscale vertical profile of hydrometeor occurrence, this dissertation expands upon their results by relating these hydrometeors to the dynamical structure of mature MCSs at the convective scale.

The goal of this chapter is to give observational insight into the dynamics and microphysics of MJO convection in the central Indian Ocean. In other regions of the world previous studies have investigated the association between the kinematic and hydrometeor structure through case analyses. While these studies provide insight into specific storms, case studies do not indicate if their results are robust features of all storms. The research presented in this chapter is, to my knowledge, the first in which dual-polarimetric radar data is used to composite multiple cases and directly show that different types of hydrometeors are organized in a repeatable and systematic fashion around the dynamical structure of mature tropical oceanic MCSs. Based on the these composites conceptual models that directly associate the kinematic and hydrometeor structures of mature, oceanic MCSs in the central Indian Ocean during the active MJO are developed. The systematic relationships demonstrated in these conceptual models are important since the hydrometeor fields derived in this dissertation are indicative of microphysical processes and their relation to storm-scale air motions. The next chapter will use these conceptual models to evaluate microphysical parameterizations and validate the interactions of dynamics and microphysics within numerical simulations.
2.2 Data / Methodology

2.2.1 S-PolKa Data and the Particle Identification Algorithm

The NCAR S-PolKa radar is a dual-wavelength (10.7 and 0.8 cm), dual-polarimetric, Doppler radar that was deployed during DYNAMO/AMIE on Addu Atoll (0.6°S, 73.1°E) in the Maldives from 1 October 2011 through 14 January 2012. It has a beam with of 0.92° and peak power of 600 kW. The radar’s scan strategy, which is detailed in Zuluaga and Houze (2013), Powell and Houze (2013), and Rowe and Houze (2014), included a series of elevation-angle scans at fixed azimuths (referred to in radar terminology as range-height indicator, or RHI, scans) that were horizontally

![Figure 2.1: Azimuthal portion of S-PolKa domain containing high resolution RHI scans within 100 km of the radar shown in gray.](image)
spaced every two degrees between azimuthal angles 4°- 82° and 114°-140°. These RHIs provided vertical cross-sections of the cloud population recorded between elevation angles of 0° and 45°. This dissertation only considers the 10.7 cm (S-band) wavelength RHI scans within 100 km of the radar (gray regions in Figure 2.1). Beyond that range the antenna’s 0.92° beam width does not provide sufficient resolution. S-PolKa’s post-experiment data processing procedures are detailed in Powell and Houze (2013) and Rowe and Houze (2014). Given that Addu Atoll is less than three meters above sea level and isolated from larger land masses, S-PolKa provides one of the first dual-polarimetric datasets of purely tropical, oceanic convection.

This chapter focuses on the eleven rainy periods identified and analyzed by Zuluaga and Houze (2013). Each of the rain events is a 48-hr period centered on a maximum in the running-mean of the 24-hr total accumulated rain observed by the S-PolKa radar. All of these rain events occurred during active periods of the MJO, when MCSs are most prevalent (e.g. Chen et al., 1996; Houze et al., 2000; Barnes and Houze, 2013; Rowe and Houze, 2014). Within the eleven rain events, individual radial velocity RHIs that display layer lifting consistent with the convective updraft and midlevel inflow trajectories shown in Figure 2.2 from Kingsmill and Houze (1999a) are identified. The rearward tilt associated with the convective updraft is a standard feature of mature convective elements that is associated with negative horizontal vorticity generated by the horizontal gradient of buoyancy at the edge of the downdraft cold pool (Rotunno et al., 1988). Thus, the convective regions identified in this dissertation are representative of the mature stage of a generic deep convective cell. As will be illustrated below, this structure is so robust that it is easily identified in single-Doppler radar data as a channel of air entering from the boundary layer, tilting upward where it converges with oppositely flowing air associated with the downdraft, and reaching a point at cell top where the flow splits as a result of cloud-top divergence. Mature stratiform regions are
characterized by a layer of a subsiding midlevel inflow that has lighter rain and a melting-layer bright band (Kingsmill and Houze, 1999a). Convective updraft and midlevel inflow layers are routinely observed in the DYNAMO/AMIE dataset and persist for long periods of time. Isolated or small convection and MCSs that have recently formed may not have these kinematic structures and are thus excluded. Mature MCSs are an important component of the MJO cloud population, especially due to their top heavy latent heating profile (Barnes et al., 2015).
RHIs have been selected based only on their radial velocity structure; they are not selected on the basis of their dual-polarimetric variables. This detail is important since, as will be explained below, the latter is composited relative to the kinematic structure seen in the radial velocity field. In order to avoid biasing results toward any one storm, only one RHI from an MCS’s convective and/or stratiform region is selected. When a storm has multiple RHIs with layer lifting, only the RHI with the most distinct airflow layer is selected. Using this criteria 25 mature convective inflow RHIs and 37 mature stratiform midlevel inflow RHIs are identified. The stratiform RHIs are further subdivided into nine stratiform RHIs with a leading-convective line and 28 stratiform RHIs without a leading-convective line, for reasons discussed below. The requirement that only one RHI be taken from each convective/stratiform region is one of the largest limitations on the size of the dataset since 5-10 RHIs from each MCS commonly were characterized by layer lifting. Kingsmill and Houze (1999a) showed that the airflow through a mature MCS is highly three-dimensional with the direction of the lower-level updraft inflow and midlevel downdraft inflow being determined by the direction of the large-scale environmental wind. Therefore, the inflow intensities may be underestimated in the composites and none of the results are based on the intensity of the inflows, only on their slopes.

S-PolKa’s alternating horizontally and vertically polarized pulses allows several dual-polarimetric radar variables that provide information about the dominant types of hydrometeors affecting MCS precipitation to be calculated. These variables include differential reflectivity (Z_{DR}), specific differential phase (K_{DP}), correlation coefficient (\rho_{hv}), and linear depolarization ratio (LDR). These variables indicate the size, shape, orientation, and water phase of the hydrometeors. Z_{DR} is sensitive to the tumbling motion, shape, density, and dielectric constant of hydrometeors. Large, oblate particles have Z_{DR} values greater than zero. Hydrometeors that are nearly spherical
and/or tumbling have $Z_{DR}$ values near zero. Because of the sensitivity of $Z_{DR}$ to the dielectric constant of the target, dry ice particles have lower values of $Z_{DR}$ than liquid water drops or water-coated ice particles of the same size. $K_{DP}$ is also sensitive to the orientation and dielectric constant of the hydrometeors, which results in large, oblate raindrops being characterized by very high $K_{DP}$ values and ice particle aggregates having slightly elevated values. $\rho_{hv}$ indicates the diversity in the size, shape, orientation, and water phase of the hydrometeor population. Most meteorological echoes are associated with an $\rho_{hv}$ of nearly one but $\rho_{hv}$ decreases to between 0.95-0.85 when the hydrometeor population becomes more diverse. LDR also indicates the hydrometeor diversity within a radar sample volume. While large negative values of LDR indicate that the hydrometeor population is uniform, small negative LDR values indicate that the hydrometeors within the radar echo volume are tumbling, oriented, and/or have different sizes, shapes, and water phases. For a comprehensive description of dual-polarimetric radar variables see Bringi and Chandrasekar (2001).

Vivekanandan et al. (1999) developed a particle identification algorithm (PID) that uses these dual-polarimetric variables and the closest rawinsonde temperature profile in a fuzzy logic algorithm to identify the type of hydrometeor that dominates the radar reflection from a given a volume of the atmosphere (referred to as the radar sample volume). The sounding data used in this dissertation to determine the temperature profile were obtained from a rawinsonde station approximately 10 km from the S-PolKa radar site and were part of the DYNAMO/AMIE sounding array described by Ciesielski et al. (2014). Each hydrometeor type classified by the PID is independently assigned an interest value between 0 and 1, which represents the likelihood of that hydrometeor being the dominant particle in that radar sample volume. The interest value is based
on the weighted sum of two-dimensional membership functions, which express the dual-polarimetric and temperature ranges associated with each type of hydrometeor. Table 2.1 provides the approximate dual-polarimetric and temperatures ranges specified by the membership functions during DYNAMO/AMIE. The hydrometeor type with the largest interest value (i.e. closest to one) is output by the PID. The hydrometeor types analyzed in this dissertation include:

Table 2.1: Approximate range of values for hydrometeor types in PID

<table>
<thead>
<tr>
<th>Hydrometeor Type</th>
<th>ZH (dBZ)</th>
<th>ZDR (dB)</th>
<th>LDR (dB)</th>
<th>KDp (° km⁻¹)</th>
<th>ρHV</th>
<th>T (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Graupel/Rimed Aggregates (G/RA)</td>
<td>30-50</td>
<td>-0.1 - 0.76</td>
<td>-25 - -20.17</td>
<td>0.08 - 1.65</td>
<td>0.89 - 0.96</td>
<td>-50 - 7</td>
</tr>
<tr>
<td>Graupel/Rain (G/R)</td>
<td>30-50</td>
<td>0.7 - 1</td>
<td>-25 - -20.17</td>
<td>0.1 - 1.7527</td>
<td>0.85 - 0.98</td>
<td>-25 - 7</td>
</tr>
<tr>
<td>Hail (H)</td>
<td>50-90</td>
<td>-3 - 1</td>
<td>-25 - -10.4</td>
<td>0 - 0.2</td>
<td>0.88 - 0.96</td>
<td>-50 - 30</td>
</tr>
<tr>
<td>Hail/Rain (H/R)</td>
<td>50-90</td>
<td>1.4 - 5</td>
<td>-27 - -25.5</td>
<td>1 - 5</td>
<td>0.86 - 0.97</td>
<td>-25 - 30</td>
</tr>
<tr>
<td>Heavy Rain (HR)</td>
<td>45-55</td>
<td>0.34 - 4.35</td>
<td>-31 - -24.5</td>
<td>0.09 - 15.55</td>
<td>0.97 - 0.99</td>
<td>1 - 40</td>
</tr>
<tr>
<td>Moderate Rain (MR)</td>
<td>35-45</td>
<td>0.01 - 3.04</td>
<td>-31 - -24.8</td>
<td>-0.01 - 2.99</td>
<td>0.97 - 0.99</td>
<td>1 - 40</td>
</tr>
<tr>
<td>Light Rain (LR)</td>
<td>10-35</td>
<td>0 - 1.8</td>
<td>-31 - -27</td>
<td>-0.02 - 0.26</td>
<td>0.97 - 0.99</td>
<td>1 - 40</td>
</tr>
<tr>
<td>Wet Aggregates (WA)</td>
<td>7-45</td>
<td>0.5 - 3</td>
<td>-26 - -17.2</td>
<td>0.1 - 1</td>
<td>0.75 - 0.98</td>
<td>-4 - 12</td>
</tr>
<tr>
<td>Dry Aggregates (DA)</td>
<td>15-33</td>
<td>0 - 1.1</td>
<td>-26 - -17.2</td>
<td>0 - 0.168</td>
<td>0.97 - -0.98</td>
<td>-50 - 1</td>
</tr>
<tr>
<td>Small Ice Crystals (SI)</td>
<td>0-15</td>
<td>0 - 0.7</td>
<td>-31 - -23.4</td>
<td>0 - 0.1</td>
<td>0.97 - 0.98</td>
<td>-50 - 1</td>
</tr>
<tr>
<td>Horizontally Oriented Ice Crystals (HI)</td>
<td>0 - 15</td>
<td>1 - 6</td>
<td>-31 - -23.4</td>
<td>0.6 - 0.8</td>
<td>0.97 - -0.98</td>
<td>-50 - 1</td>
</tr>
</tbody>
</table>
heavy rain (HR), moderate rain (MR), light rain (LR), graupel/rimed aggregates (G/RA), wet aggregates (WA), dry aggregates (DA), small ice particles (SI), and horizontally-oriented ice crystals (HI). The physical meaning of these categories will be discussed below.

While the PID is an extremely powerful tool, the algorithm has limitations. First, the PID only identifies the dominant hydrometeor type. The PID algorithm does not describe every type of particle present in the radar sample volume and the dominant hydrometeor type is not necessarily the most prevalent particle. Rather, the algorithm tends to describe the particle that is the largest, densest, and/or has the highest dielectric constant. For example, a few large aggregates will produce a return radar signal that is much stronger than the return from small ice crystals, even if the ice crystals are far more prevalent. This problem becomes more serious with distance from the radar since the size of the radar sample volume increases with range (Park et al., 2009). The impact of this limitation is explored in greater detail below. The accuracy of the PID is also limited since the theoretical associations between dual-polarimetric variables and hydrometeor types are complex and the dual-polarimetric boundaries of different hydrometeor types often overlap (Straka et al., 1999; Table 2.1). However, as Vivekanandan et al. (1999) point out, the soft boundaries in the PID allow for the fuzzy logic method to be one of the best methods to handle these complex relationships. Unfortunately, it is difficult to validate the complex relationships employed by the PID with aircraft data since it only identifies the dominant hydrometeor type. A few of the studies that conduct such a comparison are discussed below. The validity of the PID is also impacted by the quality of the radar data, which is degraded by non-uniform beam filling, attenuation, partial beam blockage, and noise. While studies such as Park et al. (2009) explicitly account for these factors in their PID algorithm, this dissertation does not. However, these radar quality issues are likely not a serious problems in this data. The S-PolKa radar experienced very little attenuation
during DYNAMO/AMIE. Beam blockage is not an issue since both RHI sectors have an unobstructed view of the ocean. S-PolKa data becomes noisy near the edges of the echo, but this only has a minor effect on the results since these areas are manually removed. Finally, the PID is limited by the accuracy of the assumed temperature profile. For example, errors in the height of the melting level can incorrectly place rain about wet aggregates. The temperature profiles have been manually edited to try to mitigate this problem.

2.2.2 Graupel and Rimed Aggregates

The process of an ice crystal collecting supercooled water droplets is called riming. Graupel is a hydrometeor that has undergone so much riming that the ice particle’s original crystalline structure is no longer distinguishable. While dual-polarimetric data identifies when riming has occurred, they do not provide a measure of the degree of riming and cannot demonstrate with certainty that a particle is sufficiently rimed to be characterized as graupel. Thus, the dual-polarimetric radar returns from a graupel particle are difficult to distinguish from those of a large aggregate of ice particles that has been affected by some riming but not enough to disguise its composition as an aggregate of ice crystals. To indicate this uncertainty, the PID includes a category called “graupel/rimed aggregates (G/RA).” This uncertainty in G/RA distinction is not an especially serious handicap since the primary goal of this dissertation is to determine where riming is likely to have been occurring.
2.2.3 Distinctiveness of the Hydrometeor Categories and Validation by Aircraft

Due to the limitations of the PID algorithm it is important to investigate the validity of the DYNAMO/AMIE PID. The Centre National d’Etudes Spatiales (CNES) Falcon aircraft was stationed on Addu Atoll from 22 November through 8 December 2011 (Yoneyama et al., 2013). Based on a few flight tracks within the S-PolKa domain, Martini et al. (2015) concluded that the PID classifications were generally accurate. However, since Martini et al. (2015) was restricted to a small dataset, this dissertation analyzed the PID’s accuracy through several more comprehensive methods. As stated above, membership functions are used in the PID to define the range of dual-polarimetric values and temperatures associated with each hydrometeor type. Table 2.1 shows that these membership functions often overlap. Thus, before using the PID it must be established that each of the eight hydrometeor types represent radar sample volumes that have unique dominant hydrometeor species. Figure 2.3 shows the observed distribution of dual-polarimetric variables of all radar sample volumes classified as a given hydrometeor type within the mature convective and stratiform regions analyzed in this chapter. The red line is the median of those radar sample volumes, the blue lines are the 25% and 75% quartiles, the black lines encompass 99.3% of the data, and the red stars are outliers. Based on Figure 2.3 it is evident that the fuzzy logic method classifies the dominant hydrometeors into groups with unique observed dual-polarimetric properties.
Figure 2.3: Bar charts showing the distribution of (a) reflectivity, (b) differential reflectivity, (c) temperature, (d) linear depolarization ratio, (e) correlation coefficient, and (f) specific differential phase of all radar volumes in convective updraft regions and contain a contiguous region of graupel/rimed aggregates, heavy rain, moderate rain, light rain, wet aggregates, dry aggregates, small ice particles, and horizontally-oriented ice crystals. The red line is the median, the blue lines are the 25% and 75% quartiles, the black lines represent 99.3% of the data, and the red stars are outliers. (g-l) Same as (a-f) except for stratiform midlevel inflow regions.
2.2.4 Examples of the PID Algorithm Choices and Associated Uncertainty

As mentioned previously, the PID assigns each hydrometeor type an interest value between zero and one to indicate the likelihood of that particle type being the dominant hydrometeor in that radar sample volume. The difference between the two largest interest values can be interpreted as a measure of certainty in the classification. Larger differences indicate that the dual-polarimetric data is consistent with the presence of only one dominant type of hydrometeor. Small differences indicate that the dual-polarimetric data is being influenced by multiple hydrometeor types. Figure 2.4 illustrates the use of the interest value difference as a way of gauging certainty in the PID’s choice of dominant hydrometeor type for a convective updraft and a stratiform midlevel inflow example. Figures 2.4a and 2.4d show the hydrometeor type with the largest interest value (1\textsuperscript{st} PID), Figures 2.4b and 2.4e show the hydrometeor type with the second largest interest value (2\textsuperscript{nd} PID), and the difference between the interest values of the 1\textsuperscript{st} and 2\textsuperscript{nd} PID is shown in Figures 2.4c and 2.4f. In order for the algorithm to classify a hydrometeor type it must have an interest value of at least 0.5. White regions in the 2\textsuperscript{nd} PID (Figure 2.4b and 2.4e) near 5 km and in the upper portions of the stratiform midlevel inflow region represent regions where only one hydrometeor type satisfies the 0.5 requirement.

Near the 5 km level, red colors in Figures 2.4c and 2.4f indicate large interest value differences and very high certainty that a single type of particle dominates the radar echo. In both the convective and stratiform cases, the radar echo in this layer is overwhelmingly dominated by WA; Figures 2.4b and 2.4e shows that no second particle type is identified. This structure marks the melting layer and is consistent with these regions containing a mixture of frozen particles that are falling and melting. In the convective example, high certainty is also seen as narrow spikes of large
Figure 2.4: Vertical cross-section showing the (a) most likely dominant hydrometeor type (1st PID), (b) second most likely dominant hydrometeor type (2nd PID), and (c) the difference in their interest values for a convective updraft at 0250 UTC on 24 October 2011. The black lines outline the convective updraft region. (d-f) same as (a-c) except for a stratiform midlevel inflow region at 0150 UTC on 18 November 2011 and the black lines outline the stratiform region.
interest value differences that occur above the 5 km and extend up to 10 km. These spikes suggest that the algorithm is certain that at least a small region of DA surrounds the G/RA particles in convective elements. However, the reduced interest value differences surrounding these spikes indicate that the full spatial extent occupied by DA is less certain.

While Figure 2.3 indicates that the hydrometeor types are statistically associated with distinct dual-polarimetric characteristics, it is important to analyze the spatial distribution of the particles since the overlapping membership functions could cause the classification in regions of small interest value differences to be somewhat random from one point to the next. This randomness does not appear to be a problem, all hydrometeors appear to be organized in a physically meaningful manner. The blue colors in Figures 2.4c and 2.4f indicate that both the convective and stratiform examples have a region of reduced PID certainty above 5 km (the approximate 0°C level) with DA as the 1st PID and SI as the 2nd PID. This result is reasonable in glaciated regions because ice hydrometeors are of a similar character but have a continuous spectrum of sizes. The DA category corresponds to larger ice particles, while the SI category corresponds to smaller ice particles. The final judgment of the PID algorithm (Figure 2.4a) in the convective example indicates that the larger DA are the dominant producer of the radar signal, which is physically reasonable since the turbulent air motions within the upper portion of the cell cause ice particles to clump into large aggregates and these large particles are more readily detected by the radar. However, the small interest value differences throughout this region (Figure 2.4c) suggests that smaller SI particles are also likely present and it is difficult to say which particle type is most numerous at a given point in time and space. Around the edges of the echo the PID expresses certainty that the SI particles dominate. This gradation from interior core to outer edge of the upper echo is another physically reasonable result of the PID in the convective echo.
The certainty associated with the frozen hydrometeors also accurately represents the inherent variability of the hydrometeor population in the stratiform region. Between 5 and 8 km, the PID’s first choice is the larger DA (Figure 2.4d) and its strong second choice is SI (Figure 2.4e and 2.4f). In the first few kilometers above the melting layer branched crystals and larger aggregates are expected (Houze and Churchill 1987; Houze 2014, p. 58-59) and their large size causes them to dominate the radar signal. Therefore, it is reasonable that the PID detects larger particles in this layer even though the 2nd PID suggests that many smaller ice particles are also likely present. In the uppermost kilometers of the stratiform echo, Figure 2.4d and 2.4e reveals that the PID algorithm is highly certain that SI are the dominant particle type, there is no viable second hydrometeor type. This zone corresponds to the red region in Figure 2.4f, which is a quantitative indication of this very high certainty. Given that vertical motions in the stratiform regions are relatively weak and cannot generate large aggregates or advect them upward, it is physically reasonable that the PID identifies only SI in the upper portions. Thus, in the glaciated regions of convective and stratiform precipitation, the 1st PID systematically describes the hydrometeor type that dominates radar signal and the 2nd PID portrays the variability of the ice hydrometeor population.

The PID algorithm is less certain in its G/RA category. These particles are seen in the convective case as four spikes extending up to 8 km in height and in the stratiform case as shallow, intermittent pockets along the top of the WA layer (Figure 2.4a and 2.4d). In both examples, these G/RA regions are characterized by small interest value differences (Figures 2.4c and 2.4f) and their second particle choice is most often the “graupel/rain (G/R)” category (Figure 2.4b and 2.4e, Table 2.1). The main distinction between the G/RA and G/R categories is that the later suggests that rimed particles are melting and mixed with liquid hydrometeors. Combining the 1st and 2nd
PID suggests that rimed particles are present and may be starting to melt, which is physically reasonable since the regions of G/R in the 2nd PID are very near the melting level at 5 km. While G/RA are expected in the convective example, the presence of G/RA in stratiform regions might seem surprising. However, it will be shown that such occurrences have been observed in previous studies and is physically reasonable.

The reduced certainty is not always an indication of the natural variability in the hydrometeor population. Evaluation of the 2nd PID on a case by case basis is important. Figures 2.4c and 2.4f show a region of small interest value differences below the melting layer at ~3-4 km altitude, which is at the top of the rain layer in both the convective and stratiform cases. This uncertainty in the PID output occurs because the 2nd PID algorithm is incorrectly identifying large raindrops as WA, as indicated by Figures 2.4b and 2.4e. At this level, the temperature in the Gan soundings (not shown) are too warm for WA to exist. Therefore, only the rain categories shown in the 1st PID are correct in this instance. Since this dissertation only uses the 1st PID in its composites, this misclassification in the 2nd PID does not impact the results.

While the PID does not comprehensively describe every type of particle present in a radar sample volume, the comparison of the PID’s first and second choices provides a high level of confidence that the algorithm’s first choice is a physically reasonable assessment of the dominate hydrometeor type. Confidence in the PID is further bolstered since even the 2nd PID is physically reasonable above the melting level and consistent with the known variability of the hydrometeor population. Similar patterns in the 1st and 2nd PID fields are found in the other convective and stratiform RHIs considered as a part of this dissertation.

In order to investigate the sensitivity of these results to the NCAR PID methodology, a different fuzzy logic based hydrometeor classification algorithm described by Dolan and Rutledge (2009)
and Dolan et al. (2013) was applied. Figures 6 and 7 in Rowe and Houze (2014) indicate that the two algorithms produce similar hydrometeor structures and supports the presence of DA and G/RA above the melting layer.

### 2.2.5 Compositing Technique

Spatial compositing of RHIs allows this dissertation to directly determine where hydrometeors occur in relation to the air motion patterns in the convective and stratiform portions of mature oceanic, tropical MCSs and assess the consistency of those relationships. Although, mature convective and stratiform regions exhibit systematic airflow patterns, these patterns do not always occur on the same horizontal scale. The relationship between the hydrometeors and airflow is most clear when each individual convective and stratiform RHI is scaled to a common size. The compositing method consists of four-steps. To illustrate these steps, consider a convective updraft layer with a contiguous region of WA identified by the PID algorithm.

1. **Identify the portion of the radial velocity RHI that contains the convective updraft layer.**

   This “convective updraft region” is determined according to the radial-velocity convergence and divergence signatures. In the horizontal direction, the region is bounded by the convergence near the surface and the width of the divergence signature aloft. The vertical extent of the convective region starts at the surface and ends where the echo becomes too weak to be detectable (typically at approximately the 5 dBZ echo contour). Figure 2.5e shows an example of a convective updraft region identified in this way. The solid black lines surround the convective element. Table 2.2 lists the width and height of
Figure 2.5: RHI scan showing (a) reflectivity, (b) hydrometeor type from the PID, (c) $Z_{DR}$, (d) $K_{DP}$, (e) radial velocity, (f) temperature, (g) $\rho_{hv}$, and (h) LDR through a convective updraft at 0250 UTC on 24 October 2011. The black line surrounds the convective updraft region. The dashed line in (b) shows the convective updraft line.
Table 2.2: Width, height, and slope of convective updraft regions

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</table>

Each convective region so identified. The width of these convective updrafts (Table 2.2) are roughly comparable to the width observed during TOGA COARE by Kingsmill and Houze (1999a) (Figure 2.2a).

2. Draw a line to approximate the center of the sloping convective updraft (dashed black line in Figure 2.5e). This line starts at the surface, where the convergent updraft layer begins to slope upward, and ends at the center of the base of the radial-velocity
divergence signature. Table 2.2 lists the slopes of all updrafts so identified. The slope of these convective updrafts (Table 2.2) are comparable to the slopes observed during TOGA COARE by Kingsmill and Houze (1999a) (Figure 2.2a).

3. **Draw a polygon around the contiguous WA region in the PID data.** The polygon is drawn manually in order to remove any noise or artifacts in the PID data and is only done if the WA are contiguous over at least two radar sample volumes. Multiple polygons are outlined if several distinct areas of contiguous WA exist within the convective updraft region.

4. **Translate and scale the location of the WA polygon to a generic convective updraft.** All convective updrafts do not slope toward the right as shown in Figure 2.2a. If a convective updraft line slopes toward the left, the mirror image of the convective updraft line and wet aggregate region is first taken. This ensures that all convective updraft lines slope toward the right. Then, the convective updraft line is stretched or compressed so that its slope exactly matches the slope of the generic convective updraft. The slope of the generic convective updraft is approximately equal to the mean updraft slope of the 25 convective RHIs. The line is also horizontally translated so that it lies exactly on top of the generic convective updraft. This process provides horizontal and vertical scaling and translation factors, which are then applied to the WA polygon to obtain its location within the generic convective updraft. This scaling process accounts for differences in the slope and horizontal extent of each convective RHI, which is important since Table 2.2 indicates that the width and slope of these convective updrafts varies substantially.

The four steps outlined above are repeated for all contiguous WA regions in all of the convective updraft RHIs, which results in a composite showing where WA are located relative to
the convective updraft as shown in Figure 2.7c. The sloping black line represents the generic convective updraft and the shading represents the frequency with which WA occur at a given location relative to the updraft. This process is repeated for each of the eight hydrometeor categories analyzed in this chapter. Since these composites only consider the location of each hydrometeor type relative to the airflow and only contain one RHI scan from each identified convective updraft region, they do not indicate the overall horizontal coverage or duration (i.e. the total amount) of the hydrometeors. These composites are used solely to explore where different types of hydrometeors are located relative to the kinematic structure of a MCS.

Stratiform midlevel inflow regions are subjected to the same compositing technique. The horizontal extent of the stratiform region is based on where the speed of the midlevel inflow exceeds the ambient radial velocity. The midlevel inflow is represented as a line that runs along the base of the accelerated flow. This line starts where the flow begins to converge and accelerate beneath the anvil and terminates where the flow returns near to its ambient speed within the stratiform precipitating region. Figure 2.6e shows the radial velocity field associated with a stratiform region on 18 November 2011. The solid black lines outline the stratiform midlevel inflow region. The dashed black line shows the midlevel inflow line. Table 2.3 lists the width and height of each stratiform region and the slope of each midlevel inflow line. The width of these midlevel inflow regions (Table 2.3) are comparable to the width observed during TOGA COARE by Kingsmill and Houze (1999a) (Figure 2.1b).
Figure 2.6: Same as Figure 2.5 except for a stratiform midlevel inflow region at 0150 UTC on 18 November 2011. The black line surrounds the stratiform region. The dashed line in (b) shows the midlevel inflow line.
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2.3 Hydrometeors in Mature Sloping Convective Updraft Channel

The radial velocity shown in Figure 2.5e demonstrates that, despite only having radial velocity data, the mature convective updraft was readily apparent. By analyzing the dual-polarimetric data the microphysical processes associated with each hydrometeor type can be understood. The case shown in Figure 2.5 was selected since its dual-polarimetric features were very distinct and representative of the other convective updraft RHIs. The reflectivity field (Figure 2.5a) showed four spikes in the 40-dBZ reflectivity above an altitude of 5 km at widths of 10-20 km. These reflectivity peaks were collocated with low $Z_{DR}$ values (Figure 2.5c), which suggests that these particles were rimed (e.g. Aydin and Seliga, 1984; Straka et al., 2000) and is consistent with the PID’s classification of G/RA (Table 2.1). The moderate reflectivity and low $Z_{DR}$ adjacent to these reflectivity spikes is indicative of aggregated ice crystals (e.g. Bader et al., 1987; Straka et al., 2000; Andrić et al., 2013) and was classified by the PID as DA. These DA were distinct from the melting particles classified as WA since WA had lower $\rho_{hv}$ and higher LDR (Figures 2.5g-h) (e.g. Zrnić et al., 1993; Straka et al., 2000; Brandes and Ikeda, 2004). Figure 2.5c shows that $Z_{DR}$ rapidly increased as particles fell below the melting layer. This dual-polarimetric signature can be interpreted as an indication of very large drops and/or heavily water-coated aggregates, which is consistent with HR and was expected given the G/RA above. The high K$_{DP}$ below an altitude of 5 km in Figure 2.5d further supported the presence of HR below the melting level (e.g. Straka et al., 2000). Thus, the microphysical and dual-polarimetric characteristics associated with each hydrometeor type were physically reasonable and consistent with previous studies.
Figure 2.7 shows the hydrometeor composites based on the 25 mature convective updraft RHIs obtained during the active stage of the MJO in the central Indian Ocean. The black line sloping upward toward the right represents the convective updraft, with flow approaching from the left, ascending in a sloping channel, and diverging at the top. The colors represent the hydrometeors’ frequency of occurrence. The color bar varies in each panel so the distribution of each hydrometer type is clearly depicted. The horizontal and vertical axes are expressed as fractions of the width and height of the convective updraft core and are referred to as the normalized width and height, respectively. Given that the spatial compositing technique distorts the vertical axis by stretching and compressing each RHI, the relationship between each frozen hydrometeor type and temperature is presented in Table 2.4. Using all radar sample volumes classified as a given hydrometeor table, Table 2.4 lists the median temperature of each frozen hydrometeor type. Additionally, Table 2.4 lists the average temperature of the coldest and warmest 10% of each hydrometeor type.

The lack of spatial spread in these composites demonstrates that hydrometeors were organized in a systematic manner with respect to the convective updraft. Below the melting layer, the updraft core was characterized by moderate rain (Figure 2.7e), which is consistent with previous observations of monsoonal squall lines in the South China Sea (Jung et al., 2012; Wang and Carey, 2005) and West African squall lines (Evaristo et al., 2010). The heaviest rain occurred just downwind of the convective core (Figure 2.7a), which is expected since the relatively large, heavy drops can only be advected short distances. While these regions of HR were likely co-located with convective-scale downdrafts, which is consistent with the sloping structure of the convective cells, this downward air motion cannot be resolved with single-Doppler radar.
Figure 2.7: Composites showing the location of (a) heavy rain, (b) light rain, (c) wet aggregates, (d) small ice crystals, (e) moderate rain, (f) graupel/rimed aggregates, (g) dry aggregates, and (h) horizontally-oriented ice in convective updrafts. The horizontal and vertical axes are expressed as fractions of the width and height of the convective updraft core and are referred to as the normalized width and height, respectively. The black line represents the composite convective updraft. Shading represents the frequency of each hydrometeor type at that location relative to the convective updraft. The color bar in each panel varies.
WA sometimes occurred at the melting level, and they were slightly more common downwind of the updraft core (Figure 2.7c). In convective regions, WA were likely associated with small-scale weak upward or downward motion (unsolved by this dataset). Thus, given the vigorous upward motion characteristic of a convective updraft core, it is not surprising that the WA were somewhat less common within the core. WA usually occurred in narrow layers but Figure 2.7c seems to suggest that WA extended over a rather deep layer. A few factors are likely influencing the depth of the WA layer in these composites. Table 2.2 indicates that the height of the convective cores varied by more than 5 km. Thus, the compositing technique, which scaled each core to the same height, distorted the vertical extent of the WA. Additionally, the median temperature of WA was reasonable at -0.98°C (Table 2.4) but temperatures associated with WA ranged from -4.9°C to 1.1°C (Table 2.4), which is larger than expected. This relatively wide range of cold minimum temperatures could have been related to beam broadening since some of the convective RHIs occurred nearly 100 km away from the radar. Thus, while the composites correctly indicated that WA was more common downwind of the updraft, the vertical extend of these particles were likely less than indicated by the composites. These factors likely influenced all of the hydrometeor composites but was most apparent in the WA composite due to their shallow depth.

G/RA occurred just behind and below the convective updraft, where heavier particles were expected to be falling (Figure 2.7f). Jung et al. (2012) suggest that the rapid fall speed of G/RA creates downdrafts beneath the convective updraft. However, these downdrafts cannot be resolved by this dataset. Table 2.4 indicates that the median temperature of G/RA was -3.7°C and the coldest 10% of G/RA were only -12.7°C on average, which suggests that the vertical extent of G/RA were often limited to a few kilometers above the melting level and is consistent with Rowe and Houze (2014). DA were the most prevalent hydrometeor above the melting level and dominated the
convective updraft (Figure 2.7g), which is similar to monsoonal squall lines in the South China Sea (Jung et al., 2012).

A relatively thin layer of SI dominated the edges of the convective region (Figure 2.7d). Note that the analysis of the 2nd PID, which is discussed above, suggested that SI were prevalent through the depth of convective clouds but only dominated the radar signal along the edges of the echo. Pockets of HI frequently were located near echo top; however, this signal might not be meaningful since $Z_{\text{DR}}$ and LDR were often noisy in these regions.

Table 2.4: Median temperature, average of temperature of coldest 10%, and average temperature of warmest 10% of frozen hydrometeors.

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<th>Dry Aggregates (DA)</th>
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2.4 Conceptual Model of Hydrometeor Occurrence Relative to Convective Region

Airflow

Based on the systematic and physically reasonable patterns of hydrometeor occurrence discussed above, a conceptual model for the spatial pattern of hydrometeors within the mature convective updraft regions of MCSs observed during the active stage of the MJO during DYNAMO/AMIE has been developed (Figure 2.8). The black lines represent the convective updraft described in Kingsmill and Houze (1999a) and the colors correspond to the different hydrometeor types. Not every hydrometeor type is present in every mature convective updraft. This conceptual diagram shows where a hydrometeor type is most likely to be located relative to the mature convective updraft morphology, if that hydrometeor type is present. Additionally, the hydrometeors depicted in this conceptual model only describe the particles that dominate the radar signal. Hydrometeors that are smaller, less dense, or have a lower dielectric constant are likely present but are not represented. The coverage of each hydrometeor type was calculated by taking the number of radar sample volumes classified as a given hydrometeor type and dividing it by the total number of sample volumes, regardless of its hydrometeor type. Each radar sample volume is 150 m wide. This coverage describes the hydrometeors’ extent in height and one horizontal dimension along the direction of the beam. This spatial coverage was averaged over all convective regions and is listed in the color bar labels in Figure 2.8. While the denominator in the coverage includes every radar sample volume in the convective updraft region, the manual analysis did not include every sample owing to smoothing and artifact removal. Thus, the coverage percentages do not add to 100%. DA accounted for over 20% of the radar sample volumes and were the most
prevalent particles in the convective region. MR covered slightly more area than HR or LR. G/RA had the smallest areal coverage at 2%. Since only one RHI was analyzed from each storm, the overall volumetric coverage of the hydrometeors in the storm cannot be determined.

The first column in Table 2.5 shows the percentage of the mature convective updraft RHIs that contained at least one contiguous region of a given hydrometeor type. MR and DA were present
in all 25 convective updraft regions. G/RA occurred 88% of the time. It is important to emphasize, however, that since the RHIs analyzed in this dissertation contained a clear, distinct convective updraft, these percentages specifically refer to convection that was either in its mature stage or just beginning to weaken, i.e. when particles classified as G/RA would have been most likely to occur. Since only the location of these hydrometeors in single RHIs was considered, this dissertation cannot comment how these frequencies varied during the lifecycle of the convective updrafts. These matters will need to be resolved by modeling and is beyond the scope of this dissertation.

Table 2.5: Frequency of occurrence of contiguous hydrometeors regions in all RHIs

<table>
<thead>
<tr>
<th></th>
<th>Convective Updraft</th>
<th>Midlevel Inflow Without a Leading Convective Line</th>
<th>Midlevel Inflow With a Leading Convective Line</th>
</tr>
</thead>
<tbody>
<tr>
<td>Graupel/Rimed Aggregates (G/RA)</td>
<td>88%</td>
<td>64%</td>
<td>66%</td>
</tr>
<tr>
<td>Heavy Rain (HR)</td>
<td>92%</td>
<td>14%</td>
<td>11%</td>
</tr>
<tr>
<td>Moderate Rain (MR)</td>
<td>100%</td>
<td>60%</td>
<td>55%</td>
</tr>
<tr>
<td>Light Rain (LR)</td>
<td>72%</td>
<td>100%</td>
<td>100%</td>
</tr>
<tr>
<td>Wet Aggregates (WA)</td>
<td>92%</td>
<td>100%</td>
<td>100%</td>
</tr>
<tr>
<td>Dry Aggregates (DA)</td>
<td>100%</td>
<td>100%</td>
<td>100%</td>
</tr>
<tr>
<td>Small Ice Crystals (SI)</td>
<td>92%</td>
<td>89%</td>
<td>100%</td>
</tr>
<tr>
<td>Horizontally-Oriented Ice Crystals (HI)</td>
<td>20%</td>
<td>53%</td>
<td>44%</td>
</tr>
</tbody>
</table>
2.5 

Hydrometeors in Mature Stratiform Midlevel Inflow Layer

Figure 2.6e shows the increased radial velocity associated with the descending midlevel inflow at an altitude of 3-5 km. While other RHIs displayed stronger midlevel inflow layers, this example was chosen since the localized radial velocity acceleration associated with the midlevel inflow is readily apparent between widths of 10-45 km and clearly demonstrates how the stratiform region is defined. Once again, the first step is to analyze the dual-polarimetric fields associated with one RHI in order to demonstrate which microphysical processes were associated with each hydrometeor type. The top of the stratiform region was characterized by low reflectivity and slightly elevated values of Z\textsubscript{DR} (Figures 2.6a and 2.6c), which is consistent with ice crystals and the PID’s SI category (Table 2.1). As the stratiform inflow descended through the layer between 5 and 10 km, reflectivity increased but Z\textsubscript{DR} decreased (Figures 2.6a and 2.6c) which is often associated with aggregation (e.g. Bader et al., 1987; Straka et al., 2000; Andrić et al., 2013) and is consistent with the PID’s classification of DA. The general transition from SI to DA likely resulted from the relatively quiescent structure of the stratiform region, which allowed ice particles to gradually settle as they slowly took on mass via vapor deposition and produced much larger particles through aggregation with each other (Houze, 1997; Bechini et al., 2013). However, while DA dominated the radar signal at altitudes between 5-10 km due to their large size, these DA were not the only frozen hydrometeors present in this layer. The analysis of the 2\textsuperscript{nd} PID, which is discussed above, suggests that SI were likely numerous above the melting level but only dominated the radar signal at high altitudes. As aggregates fell through the melting zone between 4-5 km they developed a layer of water on their exterior and began to melt at different rates, which caused an
increase in reflectivity, $Z_{DR}$, and LDR and a decrease in $\rho_{hv}$ (Figures 2.6a, 2.6c, 2.6g-h) (e.g. Zrnić et al., 1993; Straka et al., 2000; Brandes and Ikeda, 2004). All particles in the melting layer were characterized as WA (Table 2.1).

The dual-polarimetric data contained several localized features that provide details about unique microphysical processes that occurred within turbulent portions of the stratiform region. Figures 2.6a, 2.6c, 2.6g, and 2.6h, show that low reflectivity, high $Z_{DR}$, low $\rho_{hv}$, and high LDR occurred between altitudes of ~8-10 km and widths of 35-40 km. Low reflectivity indicates that these particles were relatively small and $Z_{DR}$ values of 1-2 dB indicated that these particles were horizontally oriented. While the hydrometeors were preferentially oriented in a horizontal direction, the low $\rho_{hv}$ and low LDR indicates that the tilt of each hydrometeor varied slightly. According to Table 2.1, these particles were identified as HI. To understand the microphysical processes responsible for these hydrometeors consider the radial velocity and temperature fields (Figures 2.6e-f). Note that these crystals occurred in a region where outbound and inbound radial velocities were in close proximity and temperatures were between -10°C and -20°C. The vertical wind shear likely produced turbulence and hence isolated regions of super-saturation that, in this temperature range, facilitated enhanced depositional growth of large dendritic crystals (Mason, 1971; Hobbs, 1974). This conclusion is consistent with analyses presented in Houze and Churchill (1987), Wolde and Vali (2001), Hogan et al. (2002), Andrić et al., (2013), and Bechini et al., (2013). Note that reflectivity increased, ZDR decreased, and $\rho_{hv}$ increased immediately below this dendritic growth region, which indicates that these dendrites rapidly aggregated and is consistent with DA immediately below the HI.

Another interesting, though isolated, microphysical feature indicated by the PID was the occurrence of discrete pockets of particles classified as G/RA along the top of the WA layer in
Figure 2.6b. Rowe and Houze (2014) also found this feature in their vertical PID profiles within stratiform regions. Figure 2.9 shows two sets of vertical profiles of reflectivity and $Z_{DR}$ through a portion of the RHI shown in Figure 2.6. One set of vertical profiles contains G/RA above the WA layer. The other set only contains WA. The colored dots show where each hydrometeor type is located with respect to the dual-polarimetric profiles with G/RA in green and WA in dark blue. While both reflectivity profiles have a bright band whose intensity is greater than 40 dBZ, Figures 2.9b and 2.9d show that G/RA were identified when $Z_{DR}$ was low and WA were identified when
relatively high ZDR existed. Thus, low ZDR discriminated G/RA from WA. This pattern of high reflectivity and low ZDR is indicative of riming (e.g. Aydin and Seliga, 1984; Straka et al., 2000).

The occurrence of riming within the stratiform region of MCSs is supported with in situ observations. In the central Indian Ocean, Suzuki et al. (2006) reported pictures indicative of rimed particles within stratiform regions from videosondes used during the Mirai Indian Ocean Cruise for the Study of the MJO-Convection Onset (MISMO) in 2005. Martini et al. (2015) found quasi-spherical, rimed particles within stratiform regions observed by the CNES Falcon aircraft during DYNAMO/AMIE. Additionally, precipitation image probe data from the NOAA P-3 found graupel during DYNAMO/AMIE as it descended through the melting level of a MCS to the east of the S-PolKa domain on 24 November 2011 (N. Guy, personal communication, 2014). The presence of rimed particles near the melting level in stratiform regions is not unique to the central Indian Ocean. Using data from GATE and deductive reasoning Leary and Houze (1979b) foresaw that rimed particles were likely located atop the melting layer. Evidence of rimed hydrometeors within stratiform regions has also been found in numerical simulations and observational studies in the equatorial maritime continent (Takahashi and Kuhara, 1992; Takahashi et al., 1995), West Africa (Evaristo et al., 2010; Bouniol et al., 2010), Oklahoma (Zrnić et al., 1993), Taiwan (Jung et al., 2012), and Europe (Hogan et al., 2002).

While these rimed aggregates may have been left over from collapsing deep convection (Zrnić et al., 1993; Leary and Houze, 1979b), small-scale convection embedded within the mesoscale updraft could have produced enough supercooled water for graupel to develop within the stratiform region itself (Hogan et al., 2002; Houze and Medina, 2005; Houser and Bluestein, 2011). The majority of the hydrometeors in the stratiform region gradually descend since their terminal velocity is greater than the mesoscale updraft. However, vertical wind shear along the boundary
between the midlevel inflow and mesoscale updraft can be strong enough for Kelvin-Helmholtz instability to develop and create localized regions of upward motion capable of generating supercooled water (Hogan et al., 2002; Houser and Bluestein, 2011). It is unclear which of these mechanisms was responsible for the G/RA observed in Figure 2.6 from 0150 UTC 18 November 2011. Deep convection associated with the stratiform region completely collapsed by 0100 UTC 18 November but a loop of the PID (not shown) indicates that G/RA appeared near the brightband shortly after 0030 UTC, expanded, and persisted till 0215 UTC. Kelvin-Helmholtz Instability is associated with a bulk Richardson number less than 0.25 in a stably stratified environment (Miles and Howard, 1964). The nearest sounding occurred at 0300 UTC and the bulk Richardson number evaluated between 3 and 6 km, which experienced the largest change in wind direction, was 0.25. Thus, these G/RA could have come from the deep convection that collapsed by 0100 UTC, Kelvin-Helmholtz instability, or a combination of the two.

2.6 Stratiform Regions With and Without Leading Line Structure

Stratiform midlevel inflow layers occurred within mature MCSs with two fundamentally different structural morphologies during DYNAMO/AMIE. One type of stratiform region seen frequently by S-PolKa had a very complex structure, in which convective cells entered, intensified, and collapsed within the stratiform region in the manner described by Yamada et al. (2010). These storms are referred to as stratiform without a leading-convective line and Figure 2.6 is an example taken from one such storm. However, a fundamentally different storm morphology was witnessed in late November and late December 2011 when stratiform regions were proceeded by a convective
line and both rapidly propagated rapidly toward the east as one unit. These storms are referred to as leading-line/trailing-stratiform MCSs. Rowe and Houze (2014) found that WA occurred less frequently when the stratiform region was behind a leading-convective line. Because of these different mesoscale morphologies, and the hydrometeor differences discussed by Rowe and Houze (2014), this dissertation constructs separate hydrometeor composites for mature stratiform regions with and without a leading-convective line. Table 2.3 indicates which RHIs were associated with a leading-convective line.

The composite results of the 28 RHIs of stratiform echoes without a leading convective line are shown in Figure 2.10. The black line represents the midlevel inflow geometry with the air entering beneath the anvil on the right side of each panel and descending toward the center of the storm on the left side of each panel. The horizontal and vertical axes are expressed as fractions of the width of the midlevel inflow and the height of its base. These fractions are referred to as the normalized width and height, respectively. Hydrometeors were systematically organized around the midlevel inflow. Below the melting level, the rain intensity systematically decreased toward the anvil (toward the right), as expected since the stratiform updraft gradually ascends toward the rear of the storm (Figure 2.10a, 2.10e, and 2.10b). Similar to Evaristo et al. (2010) most of the rain was light. HR was rare, but tended to occur if the midlevel inflow reached the surface, which suggests that such stratiform RHIs were slightly more convective. Above the melting layer, the hydrometeors were layered, with WA at the melting level and bands of DA and SI at sequentially higher levels (Figure 2.10c, 2.10g, and 2.10d). The layered structure is also apparent in Table 2.4, which shows a systematic decrease in the median temperature of WA, G/RA, DA, SI, and HI for mature stratiform without a leading line. The stratiform region of continental MCSs (Park et
Figure 2.10: Same as Figure 2.8 except for stratiform midlevel inflow RHI's that are not associated with a leading-convective line. The black line represents the composite stratiform midlevel inflow.
al., 2009; Bechini et al., 2013) and squall lines in Taiwan (Jung et al., 2012) and in the West Pacific (Churchill and Houze, 1987; Evaristo et al., 2010) also have a similar layered structure. The composites also captured the small-scale G/RA and HI features (Figure 2.10f and 2.10h) discussed above. The median temperature of HI was -17.6°C (Table 2.4), which is within the -10°C to -20°C temperature range that favors the dendritic growth by vapor depositions (e.g. Bechini et al. 2013). However, the average coldest HI temperatures were well below -20°C since there were a few RHIs that had HI along their echo tops. However, similar to instances of HI in the convective updraft RHIs, these echo top signatures were less certain due to increased noise.

While the composites of the mature stratiform regions with a leading-convective line only contain nine RHIs (not shown), and are be too small to allow for statistical conclusions, they are qualitatively similar to the mature stratiform regions without a leading-convective line composites. Additionally, the two types of mature stratiform regions have similar temperatures (Table 2.4) and hydrometeor frequencies (Table 2.5). There are several reasons why these basically similar structures might be expected. First, despite their morphological differences both types of stratiform are characterized by an ascending mesoscale flow at upper levels, which will result in the rain intensity systematically decreasing toward the anvil. Second, because stratiform regions are relatively quiescent in both cases, most ice particles gravitationally settle while growing by deposition and aggregation, leaving the stratiform region with its layered structure of smaller particles at higher levels and larger particles below (Houze, 1997; Bechini et al., 2013).The similarity of the hydrometeor structure in stratiform regions with and without a leading-line emphasizes that the kinematic structure of a mature MCS is a strong organization mechanism for hydrometeors.
2.7 Conceptual Model of Hydrometeor Occurrence Relative to Stratiform Region

Airflow

Figure 2.11 presents a conceptual model for the organization of hydrometeors with respect to the midlevel inflow of mature stratiform regions in MCSs. The black lines in Figure 2.11 show the kinematic structure of the mature stratiform midlevel inflow layer identified by Kingsmill and Houze (1999a). As in Figure 2.8, the colored regions in Figure 2.11 represent the location of the dominant hydrometeor and the percentages in the color bar describe the average coverage of each hydrometeor type in the mature midlevel inflow RHIs. Not every stratiform region contains every hydrometeor type. Thus, Figure 2.11 shows where a hydrometeor type is most likely to occur.

Figure 2.11: Schematic showing the location of hydrometeor types relative to the stratiform midlevel inflow. The solid black lines in the background represent the stratiform midlevel inflow. The dashed black lines indicates the melting level. The gray line represents the surface. The colored regions depict where each hydrometeor is most likely located relative to the stratiform midlevel inflow. The percentages listed in the color bar indicate the average areal coverage of each hydrometeor within RHIs taken through stratiform midlevel inflow.

53
relative to the midlevel inflow, if that hydrometeor type is present. Additionally, hydrometeors that are smaller, less dense, or have a lower dielectric constant are likely present but not represented. LR dominates below the melting level and had an average coverage of 30%. Above the melting level, SI (20% coverage) were slightly more prevalent than DA (16% coverage). HI and G/RA represented a small portion of the overall hydrometeor coverage in stratiform regions due to the fact that they occur in isolated, thin layers. WA are an important signature of the melting layer and are especially prominent in the most intense stratiform regions. Since only the location of these hydrometeors in single RHIs is considered, this dissertation cannot comment on the volumetric concentration of these hydrometeors or how the hydrometeors and their frequencies varied during the lifecycle of a stratiform region.

Table 2.5 shows the percentage of stratiform RHIs that contained at least one contiguous region of a given hydrometeor type. The differences between the two types of stratiform regions were very small, which supports only having one conceptual model for the stratiform portion of an MCS. LR, WA, and DA were always present and HR was rarely observed. This dataset contained G/RA in approximately two-thirds of the mature stratiform RHIs. However, this dissertation examined RHIs wherein the midlevel inflow is especially distinct and were more likely to have strong instability along the boundary between the descending midlevel inflow and stratiform updraft, which is precisely the situations which favor riming (Zrnić et al., 1993; Leary and Houze, 1999a; Hogan et al., 2002). Rowe and Houze (2014) considered all RHIs within the stratiform region and found that pockets of G/RA pockets occurred relatively infrequently in the overall sense. It thus appears that the well-defined midlevel inflow structure favors the occurrence of the pockets of G/RA. A similar consideration may also account for the relatively high frequency of HI (i.e. dendritic growth).
2.8 Conclusions

The S-PolKa radar was deployed as part of the DYNAMO/AMIE field campaign on Addu Atoll in the Maldives with dual-polarimetric and Doppler capabilities. Moncrieff (1992) and Kingsmill and Houze (1999a) used numerical simulations and airborne radar data, respectively, to demonstrate that air moves through mature MCSs in distinct layers. S-PolKa’s Doppler capability enables the convective updraft and stratiform midlevel inflow layers that are discussed in those papers to be identified. Additionally, since S-PolKa is a dual-polarimetric radar it is capable of remotely distinguishing different hydrometeor types, and therefore microphysical processes, over large volumes of space. This chapter capitalizes on this capability by applying the NCAR particle identification algorithm (PID) (Vivekanandan et al., 1999) to identify the most likely dominant hydrometeor within each volume of space observed by the radar. Focusing on mature MCSs observed during the active stage of the MJO, the research presented in this chapter used the radial velocity and PID fields to develop conceptual models that illustrated where hydrometeors, and their associated microphysical processes, are located with respect to the airflow patterns in MCSs.

To achieve this result, a spatial compositing technique was applied to the PID and Doppler-radar data to show where the different hydrometeor types, identified by the PID algorithm, were located with respect to the convective updraft and stratiform midlevel inflow layers in the MCSs. While every MCS does not contain every hydrometeor type, the conceptual models presented in Figures 2.8 and 2.11 show where these hydrometeors are located if they are present. These conceptual models only depict which hydrometeor type dominates the radar signal. Smaller
hydrometeors or particles with a lower dielectric constant or lower density are likely present but are not represented in the conceptual models. In mature convective updraft regions:

- The convective core is characterized by MR at low levels and DA aloft. DA are the most prevalent hydrometeor above the melting level, as expected since the turbulent upper portions of a convective cell favor aggregation.

- The heaviest rain and G/RA occur just behind and below the convective core. These are likely associated with the convective-scale downdraft, whose outflow converges near the ocean surface with the updraft inflow channel. G/RA particles occur in narrow zones that extend just a few kilometers above the melting level. This structure is expected since the updrafts of tropical oceanic convective cells are not especially intense, and these large, heavy hydrometeors rapidly melt and fall to the surface upon exiting the sloping convective updraft.

- The PID never identified hail at any level in this dataset; the updrafts over tropical oceans are not large enough to produce hail particles that could survive in temperatures above freezing.

- The melting level of a convective region is sometimes marked by a band of WA, probably produced where smaller-scale vertical motions embedded within the updraft layer are locally weaker or downward (not resolved in this dataset).

- The edges of the convective region are often characterized by decreased rain intensity and SI.

Mature stratiform regions in DYNAMO/AMIE were associated with two types of MCSs. In some mature MCSs, convection sporadically entered, intensified, and collapsed within the stratiform region. In other mature MCSs, the stratiform region was located behind a rapidly propagating
convective line. Despite these morphological differences, both types are associated with midlevel inflow layers and have the same systematic hydrometeor patterns relative to the midlevel inflow layer.

- The rain intensity systematically decreases toward the anvil. Most of the rain is light, some is moderate, but HR is rare and tends to only occur if the midlevel inflow reaches the surface. These patterns are expected since the stratiform updraft is unable to suspend heavier particles or allow large, heavy particles to be advected over large horizontal distances.

- Ice hydrometeors occur in well-defined layers, with a vertically thin but robust band of WA at the melting level, a thick layer of DA above the melting zone, and finally SI at the highest levels. This general layering is expected since most of the stratiform region has weak vertical air motion, allowing particles to gravitationally settle (Houze, 1997).

- Pockets of G/RA occur intermittently with small-scale spatial variability just above the WA layer. Stratiform regions may contain rimed particles as a result of collapsing deep convective cores (Houze, 1997) or small, localized convection embedded within the mesoscale stratiform updraft that is associated with internal Kelvin-Helmholtz instability (Hogan et al., 2002; Houze and Medina, 2005). It cannot determine which of these riming mechanisms is acting in these RHIs. However, comparison with the overall dataset of DYNAMO/AMIE (presented in Rowe and Houze (2014)) suggests that these pockets of riming preferentially occur when the stratiform midlevel inflow layer is most robust.

- Small regions of HI are occasionally embedded within the stratiform region and are likely associated with enhanced dendritic growth by vapor deposition. Comparison with the
overall dataset of DYNAMO/AMIE (presented in Rowe and Houze (2014)) indicates that this zone is most prominent when the stratiform midlevel inflow layer is robust.

The systematic hydrometeor structure within stratiform regions, regardless of whether or not the MCS is part of a leading-line/trailing stratiform MCS, is expected. Stratiform regions may occur as a result of convection either dying out to form and/or become part of a pre-existing stratiform region or being sheared off of a leading line (Houze 2014, Chapter 6). In either case, the stratiform cloud is composed of old convective material whose air is still somewhat buoyant and contains particles formed in active convection. The ice particles continue to grow in the stratiform region by vapor deposition as a result of the weak residual buoyancy of the previously convective parent air parcels. Thus, the origins of stratiform precipitation are similar despite the nature of the arrangement of cells. Accordingly, there is no a priori reason to expect the microphysics of the stratiform regions to depend on the presence or absence of a leading convective line. This similarity implies that the structure of latent heating is fundamentally similar in all stratiform regions. Further research is necessary, but this results suggests that only one latent heating parameterization is needed to adequately describe all stratiform regions.

Several of features of the hydrometeor structure presented in this chapter have been observed in previous studies. For example, moderate rain is commonly observed within the convective updraft (e.g. Jung et al., 2012; Wang and Carey; 2005; Evaristo et al., 2010), frozen hydrometeors routinely have a layered structure in stratiform regions (e.g. Houze and Churchill, 1987, Park et al., 2009; Bechini et al., 2013; Jung et al., 2012), and graupel is occasionally observed along the melting level of mature stratiform regions (e.g. Takahashi and Kuhara, 1992; Hogan et al., 2002; Evaristo et al., 2010; Rowe and Houze, 2014, Martini et al., 2015). Given that the kinematic structure of mature MCSs is similar throughout the globe (e.g. Zipser, 1977; Keenan and Carbone,
MCSs in other geographic regions are expected to have a somewhat similar hydrometeor structure. However, some important hydrometeor differences exist between continental and tropical MCSs due to differences in the thermodynamic profile, aerosol content, and convective intensity likely (e.g. Zipser and LeMone, 1980; LeMone and Zipser, 1980). For example, it is known that tropical West African MCSs have more graupel and possibly hail, probably because of the greater buoyancy and stronger updrafts generated over land (Evaristo et al., 2010; Cetrone and Houze, 2011; Yuan et al., 2011). Additionally, Houze (1989) showed that the convective regions of mesoscale systems have different vertical velocity profiles in different tropical oceanic regions. Thus, this chapter describes the hydrometeor organization in MCSs in the central Indian Ocean but is not necessarily representative of all the MCSs observed globally.

While this dissertation focuses on mature MCSs during the active stage of the MJO during DYNAMO/AMIE, five convective and two midlevel inflow RHIs during the suppressed stage were identified and the same analysis was conducted (not shown). While the representativeness of these suppressed composites is limited by their small sample size, they are qualitatively similar to the active stage composites discussed above. This similarity between the active and suppressed stage of the MJO is consistent with Rowe and Houze (2014). These results suggest that the MJO provides the large-scale environment that favors the development of MCSs but does not fundamental change the hydrometeor structure of MCSs. Thus, the kinematic structure of a MCS is a strong hydrometeor organization mechanism.

Other studies have penetrated more deeply into the microphysical structure of convection by relating dual-polarimetric observations to combinations of microphysical and electromagnetic models (Kumjian and Ryzhkov, 2010, 2012; Kumjian et al., 2012a, 2012b, Andrić et al., 2013). Even though this dissertation uses the methodology of Vivekanandan et al. (1999) to analyze the
hydrometeor structure of mature MCSs, these hydrometeor classifications can be interpreted in terms of microphysical processes. For example, G/RA are associated with riming, DA are associated with ice particle aggregation, and WA represent melting particles. The ability to interpret the PID in terms of microphysical processes allows the conceptual models shown in Figures 2.8 and 2.11 to be used to validate numerical simulations using generally available bin and bulk microphysical schemes in high-resolution regional and cloud-resolving models. While the conceptual models presented in this chapter are not able to verify the amount and time tendency of different hydrometeor types, these conceptual diagrams will be able to verify whether or not the spatial organization of microphysical processes are accurate in relation to the mesoscale air motion patterns.

The first step in conducting such a spatial comparison is to associate each of the eight PID hydrometeor types with the hydrometeor types or microphysical processes used in numerical simulations. The association between the PID and simulated hydrometeor types is coarse. Assuming that the hydrometeors classified by the PID are precipitating, the PID categories cannot be used to verify simulated cloud water or ice, which are usually considered to be non-precipitating. HI, SI, and DA are all represented as snow in numerical simulations. G/RA and graupel in the simulation are somewhat comparable. Given that the simulated hydrometeors instantaneously melt upon reaching the melting level, the simulated 0°C level should be at the same height as the top of the WA layer. Simulated rain represents WA, HR, MR, and LR. The association of PID and model hydrometeor categories is made even more complicated since the hydrometeor definitions used in numerical simulations and the PID are fundamentally different. A more effective way to compare PID data to numerical simulations is in terms of microphysical processes. Then, DA represents simulated aggregation, G/RA are comparable to simulated riming,
and WA represents the simulated melting. These processes are calculate within microphysical parameterizations and can easily be fields of these variables can be easily output from numerical simulations. However, it is important to acknowledge that the comparisons stated above are only a general reference, each microphysical parameterization is different and needs to be individually evaluated to ensure that the proper hydrometeors and microphysical processes are being compared. Numerical simulation/radar comparisons are the subject of the next chapter.
CHAPTER 3

COMPARISON OF OBSERVED AND SIMULATED SPATIAL PATTERNS OF ICE MICROPHYSICAL PROCESSES IN TROPICAL OCEANIC MESOSCALE CONVECTIVE SYSTEMS

This chapter equitably compares the spatial pattern of ice microphysical processes produced by three microphysical parameterizations with each other, observations, and previous observations-based studies. Simulations of tropical oceanic mesoscale convective systems (MCSs) in the Weather Research and Forecasting (WRF) model are forced to develop the same mesoscale circulations as observations by assimilating Doppler-radar observations of radial velocity. The same general layering of ice microphysical processes was found in observations and simulations with deposition anywhere above the 0°C level, aggregation at and above the 0°C level, melting at and below the 0°C level, and riming near the 0°C level. Thus, the layered ice microphysical pattern portrayed in this chapter is consistent with previous conceptual models and dual-polarization radar data. Spatial variability of riming in the simulations suggest that riming in the midlevel inflow is related to convective-scale vertical velocity perturbations. Finally, this chapter sheds light on limitations of current generally available bulk microphysical parameterizations. In each parameterization, the layers in which aggregation and riming took place were generally too thick and the frequency of riming was generally too high. Additionally, none of the parameterizations produced similar details in every microphysical spatial pattern. Discrepancies in the patterns of
microphysical processes between parameterizations likely factor into creating substantial differences in the model reflectivity patterns. It is concluded that improving the parameterization of ice-phase microphysics will be essential to obtain reliable, consistent model simulations of tropical oceanic MCSs.

Publication Reference:


3.1 Introduction

The spatial distribution of microphysical processes within a convective cloud system indicates how the microscale interactions among hydrometeors are affecting the entire dynamical and thermodynamical system. Studies such as Chen and Cotton (1988) have demonstrated that accurate mesoscale simulations require accurate representations of microphysical processes, latent heating, and radiative transfer and their interactions. Latent heat absorbed or emitted during cloud microphysical processes modifies buoyancy, which, in turn, contributes to the development and maintenance of vertical air motion (e.g. Szeto et al., 1988; Tao et al., 1995; Adams-Selin et al., 2013). Ice-phase microphysical processes are essential to the development of both the convective and stratiform components of mesoscale convective systems (MCSs) (e.g. Chen and Cotton, 1988; Tao et al., 1991; Zipser, 2003). Once stratiform precipitation is formed, ice microphysical processes modify radiative heating, which can increase instability, cause turbulence, and extend the lifetime of stratiform precipitation and its associated anvil cloud (e.g. Webster and Stephens, 1980; Chen and Cotton, 1988; Churchill and Houze, 1991; Tao et al., 1996).

The impact of microphysical processes is not limited to the convective cloud systems in which they occur. The modification of the diabatic heating structure by microphysical processes influences dynamics at the global-scale. This evolving diabatic heating profile alters the global circulation through teleconnections (e.g. Hartmann et al., 1984; Schumacher et al., 2004). For example, Barnes et al. (2015) demonstrated that the latent heating profile systematically changes during the Madden-Julian Oscillation (MJO). This evolving latent heating profile is likely one reason why studies such as Vitart and Molteni (2010) have demonstrated that extratropical weather patterns are correlated to the MJO.
Carefully collected and processed dual-polarization radar data provides some of the most comprehensive data on microphysical processes in mesoscale convective cloud systems. One approach of using this data to obtain insight into the microphysical processes in convection is to use a particle identification algorithm (PID). Traditionally, PIDs have been used to identify the dominant hydrometeor type within convection (e.g. Chapter 2, Hendry and Antar, 1984; Vivekanadan et al., 1999; Straka et al., 2000; Thompson et al., 2014; Grazioli et al., 2015; Kouketsu et al., 2015). Because the PID classification is based on physical characteristics of the particles, not their exact size or density, PID classifications do not indicate hydrometeor mixing ratios. However, the particle type indicated by the PID is a good indicator of the microphysical processes that have produced the particles (i.e. deposition, aggregation, or riming). This chapter capitalizes on this capability of the PID methodology.

PID classifications have restrictions that must be considered when the data is analyzed. First, radars statistically sample large volumes that contain many particles and processes but the PID only provides data on the most likely dominant microphysical process. Additionally, the PID categories are somewhat uncertain due to overlapping classification boundaries. Nonetheless, the PID provides the most comprehensive high-resolution three-dimensional mapping of the spatial distribution of microphysical processes in convection. Ground instruments and aircraft probes provide in situ observations (e.g. Baumgardner et al., 2011), but their spatial coverage is limited. Additionally, previous aircraft probes biased results by shattering ice crystals (Korolev et al., 2011). Satellites can retrieve details about the microphysical structure over large areas and long temporal periods (e.g. Comstock et al., 2007), but the resolution is coarse. PIDs therefore remain the best option for comprehensive mapping of microphysical processes in convection.
The microphysical processes at a given location are closely linked to the evolution of the airflow. Mesoscale convective systems (MCSs), which are broadly defined as cloud systems whose contiguous precipitation span at least 100 km in any one direction (Houze, 2004; 2014), are ideal for investigating the spatial microphysical patterns in convection because they have a well-known, repeatable kinematic structure (Kingsmill and Houze, 1999a). MCSs are generally composed of two parts: a convective and stratiform region. The convective region is composed of intensely precipitating cores of relatively small horizontal scale. In the convective region, the air steeply rises as a layer originating in the lower troposphere. The stratiform region is a more expansive region of weaker precipitation characterized by a current of air that starts at midlevels within the anvil and extends toward the center of the storm as it gently subsides. This current of air is referred to as the midlevel inflow. Vertical velocities in the stratiform zone are weak, often an order of magnitude less than the horizontal wind speeds, and are characterized by broad regions of weak ascent above the midlevel inflow and weak descent below the midlevel inflow. Differences in the large-scale environment cause the horizontal morphology of MCSs to vary (Tollerud and Esbensen, 1985). Often in environments with strong lower tropospheric vertical wind shear MCSs take the form of a leading convective line with a trailing stratiform region. These MCSs are called squall lines. When MCSs develop in environments with weak low-level vertical wind shear they tend to be more amorphous with the convective regions lying alongside the stratiform region in various patterns. Despite these morphological differences all MCSs have the same convective updraft and midlevel inflow structure (Kingsmill and Houze, 1999a). Because of this repeatable kinematic structure, it is possible to composite the pattern of microphysical processes relative to the mean rising current in the convective region and the mesoscale midlevel inflow dominating the stratiform region.
Chapter 2 took this compositing approach and applied it to a PID developed at the National Center for Atmospheric Research (NCAR) to data obtained using the NCAR S-PolKa radar during the Dynamics of the Madden-Julian Oscillation / ARM MJO Investigation Experiment (DYNAMO/AMIE) (Yoneyama et al., 2013). Compositing the PID data from 36 MCSs with a midlevel inflow, Chapter 2 showed that frozen hydrometeors had a systematic layered pattern with small ice crystals near echo top, dry aggregates at midlevels above the 0°C level, and melting near the 0°C level. Additionally, graupel/rimed aggregates were found occasionally in isolated, shallow pockets just above the 0°C level. These results were insensitive to the morphology of the MCS. The same hydrometeor pattern was observed in both squall and non-squall line MCSs. While previous studies found a similar layered structure to the frozen hydrometeors and identified graupel in stratiform precipitation in individual case studies (e.g. Leary and Houze, 1979; Houze and Churchill, 1987; Takahashi and Kuhara, 1993; Hogan et al., 2002; Park et al., 2009; Evaristo et al., 2010; Jung et al., 2012; Bechini et al., 2013; Martini et al., 2015), the PID-based composite analysis conducted in Chapter 2 was the first to demonstrate that these hydrometeor patterns are statistically robust features of tropical, oceanic MCSs and have a systematic relationship to the characteristic mesoscale midlevel inflow.

If the PID results from Chapter 2 are interpreted in terms of ice microphysical processes it indicates that the dominant ice microphysical processes in MCSs over tropical oceans systematically transition downward from primarily deposition at high levels to a layer of considerable aggregation before melting, with riming occasionally occurring in shallow pockets just above the melting level. This layered pattern is consistent with the conclusions and conceptual diagrams of Leary and Houze (1979), Houze (1981; 1989), and Houze and Churchill (1987). Given
that Chapter 2 has demonstrated that this ice microphysical pattern is statistically robust, the question arises as to whether simulated microphysical processes exhibit a similar pattern.

In recent decades the number of microphysical parameterizations available has substantially increased, which has led to a large number of studies focused on comparing these schemes in a variety of storm types, atmospheric environments, and model frameworks, including tropical MCSs (e.g. Blossey et al., 2007; Wang et al., 2009; Varble et al., 2011; Van Weverberg et al., 2013; Hagos et al., 2014; Roh and Satoh, 2014). One aspect of microphysical parameterizations that has been largely neglected in prior intercomparison studies is the spatial distribution of microphysical processes. Caniaux et al. (1994) is one of the only studies that explicitly shows the spatial pattern of microphysical processes within simulated convection. However, their study was conducted using a two-dimensional anelastic model. Somewhat surprisingly, it is relatively unknown how these processes arrange themselves in three-dimensional full-physics simulations. Several studies including Donner et al. (2001) have shown the spatial pattern of latent heating attributed to specific microphysical processes. While these studies provide an indication of the spatial pattern of microphysical processes that change the phase of water, this technique provides no insight into processes that do not impact latent heating, such as aggregation. A common way to analyze the simulated spatial pattern of microphysical processes is to apply a dual-polarization radar simulator to the model output and compare the simulator’s output with observations (e.g. Jung et al., 2012; Putnam et al., 2014; Brown et al., 2015; Jung et al., 2016). While these studies provide insight into the simulated microphysical structure, it is unclear whether observed differences are attributable to the model, the simulator, or a combination of both. Knowledge of how microphysical parameterizations spatially organize microphysical processes is essentially since Roh and Satoh (2014) indicated that simulated vertical velocities and temperatures are
sensitive to the microphysical parameterization selected. The spatial distribution of ice microphysical processes is particularly important since Varble et al. (2011) demonstrated that mixed and ice phase regions in observations and simulations substantially differ and Wang et al. (2009) suggested that inconsistencies among observations and simulations result from differing treatments of frozen hydrometeors in parameterizations.

This chapter assimilates radial velocity data obtained during DYNAMO/AMIE into the Weather Research and Forecasting (WRF) model and outputs fields that show the spatial pattern of deposition, aggregation, riming, and melting. This technique constrains naturally evolving MCSs in a full-physics three-dimensional model to have the same kinematic structure as observed MCSs, while allowing the microphysical processes to interact with these circulations. This method enables the spatial pattern of simulated and observed microphysical processes to be directly and equitably compared with minimal contamination from dynamical differences. Using these model results, this chapter will explicitly investigate if three different ice microphysical parameterization schemes produce large-scale spatial patterns of deposition, aggregation, riming, and melting that are:

1. Consistent with the robust systematic ice microphysical pattern observed in the DYNAMO/AMIE PID data,
2. Consistent with our theoretical understanding of microphysics and their interaction with the dynamical structure of convection, and
3. Consistent with each other.
3.2 Methodology

3.2.1 Microphysical Interpretation of Particle Identification (PID) Algorithm

By emitting and receiving horizontally and vertically polarized pulses, dual-polarized radars obtain moments of the particle distribution in the radar sampling volume that indicate the physical characteristics of the particles. The variables obtained in this way include differential reflectivity ($Z_{DR}$), linear depolarization ratio (LDR), correlation coefficient ($\rho_{hv}$), and specific differential phase ($K_{DP}$). (See Section 2.2.1 for a brief summary of dual-polarimetric radar variables or Bringi and Chandrasekar (2001) for a comprehensive description.) Vivekanandan et al. (1999) developed a technique that classifies regions within convection by combining the dual-polarization radar variables and rawinsonde temperature profiles. Included in this frequently used algorithm, which is referred to as a particle identification algorithm (PID), are four frozen hydrometeor categories. Chapter 2 referred to these frozen hydrometeor categories as small ice crystals, dry aggregates, wet aggregates, and graupel/rimed aggregates. (For additional details about the PID used during DYNAMO/AMIE see Section 2.2.1 and Rowe and Houze (2014).) While these categories are named in terms of particle type, the names imply the microphysical processes producing the particles. Table 3.1 shows the dual-polarimetric and temperature thresholds used to define four of the frozen categories identified by the PID. Small ice crystals are associated with the smallest horizontal reflectivity and have large negative values of LDR, which suggests that these are the smallest frozen hydrometeors identified by the radar and have no preferred shape. The most likely way that a frozen particle can grow large enough to be detected by the radar and yet have such
random shape characteristics is through deposition. These small ice particles could have been advected from another part of the storm, but even in that case any subsequent growth on these small advected particles in this temperature range would most likely result from deposition. Compared to the small ice crystals, dry aggregates have higher reflectivity and their LDR values are less negative. Thus, dry aggregates are larger and more uniform in shape than small ice crystals, suggesting that dry aggregates have undergone aggregation (e.g. Bader et al., 1987; Straka et al., 2000; Andrić et al., 2013). They are referred to as dry aggregates because they are generally found well above the 0°C level, where they would not have any liquid attribute as would melting aggregates. In a region of upward motion, such as occurs above the midlevel inflow layer of an MCS, these dry aggregates would also be expected to be increasing in mass by vapor deposition, as suggested by Houze (1989). Some of the distinguishing characteristics of wet aggregates include reduced correlation coefficient and high differential reflectivity, which are indications of mixed phase particles and melting (e.g. Zrnić et al., 1993; Straka et al., 2000; Brandes and Ikeda, 2004). Previous studies have demonstrated that the combination of high reflectivity and low differential reflectivity assigned to the graupel/rimed aggregate category is predominantly associated with
riming (e.g. Section 2.5, Aydin and Seliga, 1984; Straka et al., 2000). Rimed particles, such as graupel, would also be growing by vapor deposition in in-cloud regions of upward motion. Thus, PID categorization can be used to map the spatial pattern of ice microphysical processes within a region of radar echo in an MCS: small ice crystals indicate primarily deposition; dry aggregates represent aggregation combined with deposition; graupel/rimed aggregates indicate riming possibly mixed with deposition; and wet aggregates exist where melting is occurring.

While the PID is a powerful technique for mapping the locations of different microphysical processes, it has limitations. First, the PID identifies only the process creating the hydrometeor with the largest radar return, even though multiple processes are likely occurring in a given volume of air. As a result, PID results are biased toward processes that create particles that are the largest, densest, or have the highest dielectric constant. Consequently, the process identified may not be the most prevalent process. This problem becomes more serious with distance from the radar due to the broadening of the radar beam (Park et al., 2009). The certainty of the PID is also questioned because the theoretical associations between the dual-polarimetric variables and categories are complex and involve overlapping boundaries (Tables 2.1 and 3.1) and because few validation studies exist. Vivekanandan et al. (1999) pointed out that the complex relationships used in the PID are unavoidable and the soft boundaries used in their fuzzy logic algorithm provides one of the best ways to handle them.

This chapter does not attempt to address all the uncertainties in PID algorithms. This chapter is applying the technique only to a limited type of convection, namely, MCSs observed in a tropical, oceanic environment, specifically in DYNAMO/AMIE. In the case of this dataset, Chapter 2 and Martini et al. (2015) have analyzed the performance of the PID and concluded that it was accurate. Thus, there is high confidence that the DYNAMO/AMIE PID dataset describes
the dominant ice microphysical process with sufficient accuracy for the purposes of this dissertation.

The aim of this chapter is to interpret the PID in terms of microphysical processes in a way that can determine if the spatial pattern of microphysical processes in numerical simulations is consistent with that seen in the DYNAMO/AMIE dataset. The PID cannot validate numerical simulations when it is interpreted in terms of hydrometeors because the defining characteristics of hydrometeors in the PID and numerical simulations are define hydrometeors fundamentally different. In numerical simulations, hydrometeors are defined based on the particle size and density. The PID defines hydrometeors based on their relative shape, their water phase, and the uniformity of the particles. Models compute mixing ratios of particle types; PID methods cannot determine mixing ratios. However, the PID data can be interpreted in terms of the microphysical processes producing the particles, and numerical simulations directly calculate the same processes. The comparisons in this chapter are therefore based on processes, not mixing ratios.

### 3.2.2 Classification of Microphysical Processes in WRF

Microphysical parameterizations within WRF only routinely output the hydrometeor mixing ratios, and also number concentrations if the scheme is double-moment. However, dozens of variables describe the interactions among the mixing ratios of individual hydrometeor types. By grouping these hydrometeor interaction variables in terms of the microphysical process they describe, additional three-dimensional fields depicting the production rate of microphysical processes can be output. This dissertation considers four ice microphysical processes: deposition,
aggregation, riming, and melting. The following definitions were used to group the hydrometeor interaction variables:

- Deposition included any hydrometeor interaction variable that described how a frozen hydrometeor collected water vapor.
- Aggregation included any hydrometeor interaction variable that described the collection of a frozen hydrometeor on to another frozen hydrometeor. This included any variable that described the interaction among snow, ice, graupel, and/or hail.
- Riming included any hydrometeor interaction variable that described how a frozen hydrometeor collected a liquid hydrometeor and remained classified as frozen. It is possible that a thin layer of liquid developed along the edge of the hydrometeor even though its core remained frozen.
- Melting included any hydrometeor interaction variable that described how a frozen hydrometeor became a liquid hydrometeor.

This dissertation tests three routinely available microphysical parameterizations in WRF: the Milbrandt-Yau Double Moment Scheme (MY) (Milbrandt and Yau, 2005a; b), the Morrison 2-Moment Scheme (MOR) (Morrison et al., 2009), and the WRF Double-Moment 6-Class Scheme (WD) (Lim and Hong, 2010). While MY is fully-double moment, the MOR and WD are only partially double-moment. MOR is double moment in rain, ice, snow, and graupel. WD scheme is double moment in cloud water and rain. Table 3.2 shows how the hydrometeor interaction variables in each microphysical parameterization were grouped into microphysical processes.
Table 3.2: The definition of the ice microphysical processes and the variables from each parameterization used to define each process.

<table>
<thead>
<tr>
<th>Parameterizations</th>
<th>Ice Microphysical Processes</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Aggregation</td>
</tr>
<tr>
<td>Frozen hydrometeors collecting other frozen hydrometeors</td>
<td>Frozen hydrometeors collecting liquid hydrometeors</td>
</tr>
<tr>
<td>Milbrandt – Yau (MY)</td>
<td>QCLis, QCLig, QCLsh, QCNis2, QCLih</td>
</tr>
<tr>
<td>Morrison (MOR)</td>
<td>prai, prci</td>
</tr>
<tr>
<td>WDM6 (WD)</td>
<td>Psaci, Pgaici, Psaut, Pgaacs, Pgaaut</td>
</tr>
</tbody>
</table>

3.2.3 Data Assimilation

One of the most challenging aspects of comparing observed and simulated cloud microphysical processes is their sensitivity to the dynamics. If the observed and simulated dynamical structure differ it is impossible to discern if differences in the microphysical processes are related to the processes themselves or dynamical differences. Shipway and Hill (2012) addressed this complication by developing a one-dimensional kinematic driver model that prescribed the flow and did not allow interaction among the dynamics and microphysics. While their technique
provides a straightforward method to equitably compare parameterizations, it does not address the fact that the microphysics evolve in concert with the dynamics of convection.

This dissertation uses data assimilation to require simulations to develop a mesoscale circulation similar to observations. By so doing, the microphysical processes are allowed to evolve freely but only in concert with a dynamically accurate simulation. The microphysics can then be compared with dual polarization radar observations with minimal contamination from wrong dynamics. Previous studies suggest that this technique is a valid way to investigate microphysical variability among parameterizations. Wheatley et al. (2014) assimilated radar data using a WRF-based ensemble Kalman filter and demonstrated that differences among microphysical schemes were the same whether or not radar data was assimilated. Assimilating radial velocity data and mesonet observations, Marquis et al. (2014) demonstrated that a WRF-based ensemble Kalman Filter (EnKF) could simulate supercell thunderstorms whose kinematic structure was consistent with observations. While these studies were investing a type of convection very different from tropical oceanic convection, their results suggest our approach is reasonable. However, data assimilation in Marquis et al. (2014) only accounted for the mesoscale circulation. While the mesoscale circulation is the dominant dynamical feature of MCSs, convective-scale dynamical differences exist within our simulations and likely impact some of our results. The impact of these differences are relatively minor and will be discussed where applicable.

Radial velocity is the most accurate radar variable. It involves no assumptions about the nature of the particles, and it is a direct measurement. All other radar variables are derived. Additionally, all other radar variables measure characteristics of the particle population and assimilating them could manipulate the simulated microphysical pattern. This manipulation would obscure differences among observations and simulations. Thus, radial velocity alone is the most
appropriate radar variable to assimilate for this dissertation since it forces simulations to have a similar mesoscale dynamical evolution while impacting the microphysical processes as little as possible.

Radial velocity data used in this dissertation was obtained by the NCAR S-PolKa radar during the DYNAMO/AMIE field campaign, which was located on Addu Atoll in the Maldives from November 2011 through January 2012 (Yoneyama et al., 2013). The NCAR S-PolKa radar is a dual-wavelength (10.7 and 0.8 cm), dual-polarimetric, Doppler radar that has a beam width of 0.92° and maximum range of 150 km. In order to map the horizontal and vertical distribution of convection, S-PolKa’s scan strategy during DYNAMO/AMIE consisted of a set of horizontal surveillance scans (called plain position indicator (PPI) scans) and vertical scans along specified azimuthal angles (called range height indicator (RHI) scans) repeated every 15 m. (For more details about the scan strategy used during DYNAMO/AMIE see Zuluaga and Houze (2013), Powell and Houze (2013), and Rowe and Houze (2014).) This chapter only considers PPI scans from the 10.7 cm (S-band) wavelength radar. Prior to assimilation the radial velocity data was quality-controlled and thinned. Quality-control measures included removing data in pixels that contained low signal-to-noise ratios, clutter, and/or high spectral width. Additionally, data in pixels with reflectivity less than 3 dBZ were excluded. The quality-control of the radial velocity data was conducted with the assistance of a Matlab-based quality-control package being developed by the NCAR Earth Observing Laboratory (Scott Ellis, NCAR EOL, personal communication). Given that the NCAR S-PolKa radar data had a much finer horizontal resolution (150 m) than our simulations (3 and 1 km), the radial velocity data was thinned prior to assimilation onto a 2° x 1 km grid using the super-observation technique discussed in Zhang et al. (2009).
This dissertation uses the Advanced Research version of the Weather Research and Forecasting (WRF-ARW) model version 3.5 (Skamarock et al., 2008). Details about the WRF architecture used in this dissertation is provided in Table 3.3 and the nested domains with 3 and 1 km resolution, respectively, are shown in Figure 3.1. Each of the domains were two-way nested and had 39 unevenly spaced vertical levels between the surface and 26 km with maximum resolution near the surface. Analysis was only conducted in the inner domain with 1 km resolution. The inner domain was located mostly east of the radar because the scans at low elevation angles to the west of the radar were obscured by ground clutter. The high computation expense of data assimilation prevented finer horizontal resolutions from being used. The model time step was 18 s and data was output every 15 mins. This time step and output period was appropriate given that this dissertation is only interested in the spatial pattern of the microphysical processes, not on temporal evolution of the microphysical processes.

Figure 3.1: Outer (blue square) and inner (red square) domains used in WRF simulations. The resolution of each domain is listed in parenthesis. The S-PolKa radar is located at the black dot and its domain is outlined by the black circle.
Table 3.3: WRF and EnKF setup for simulations.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>December Simulation</th>
<th>October Simulation</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>WRF Setup</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Simulation Time</td>
<td>23 December 2011; 1200-2000 UTC</td>
<td>16 October 2011; 0600-1800 UTC</td>
<td></td>
</tr>
<tr>
<td>Time Step</td>
<td>18 seconds</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vertical Levels</td>
<td>39, top at 26 km</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Domain Horizontal Resolution</td>
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<td></td>
</tr>
<tr>
<td>Nesting</td>
<td>Two-way</td>
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<td></td>
</tr>
<tr>
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<td>University of Washington (TKE) Boundary Layer Scheme</td>
<td>Bretherton and Park (2009)</td>
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<td>Longwave Radiation Parameterization</td>
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<td>Mlawer et al. (1997)</td>
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<tr>
<td>Shortwave Radiation Parameterization</td>
<td>Dudhia Shortwave Scheme</td>
<td>Dudhia (1989)</td>
<td></td>
</tr>
<tr>
<td>Surface Layer Parameterization</td>
<td>MM5 Similarity Scheme</td>
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<td></td>
</tr>
<tr>
<td>Land Surface Parameterization</td>
<td>United Noah Land Surface Model</td>
<td>Tewari et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>Microphysics Parameterizations</td>
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<td>Milbrandt and Yau (2005a; b)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Morrison 2 - Moment Scheme</td>
<td>Morrison et al. (2009)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>WRF Double - Moment 6 - Class Scheme</td>
<td>Lim and Hong (2010)</td>
<td></td>
</tr>
<tr>
<td><strong>EnKF Setup</strong></td>
<td></td>
<td></td>
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<tr>
<td>First Assimilation Time</td>
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<td>1200 UTC</td>
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<tr>
<td>Assimilation Interval</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Assimilated Data</td>
<td>S-PolKa Radial Velocity</td>
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<td></td>
</tr>
<tr>
<td>Number of Members</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Ensemble Initiation and Boundary Conditions</td>
<td>ERA-interim perturbed using WRFDA cv option 5</td>
<td>Barker et al. (2005)</td>
<td></td>
</tr>
<tr>
<td>HROI</td>
<td>90 km, 30 km, Gaspari and Cohn 5th Order</td>
<td>Gaspari and Cohn (1999)</td>
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<td>VROI</td>
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<tr>
<td>Relaxation Coefficient</td>
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<td></td>
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<tr>
<td>Perturbed Variables</td>
<td>perturbation potential temperature, zonal &amp; meridional wind, cloud water mixing ratio, vapor mixing ratio, rain mixing ratio, base-state &amp; perturbation geopotential, base-state &amp; perturbation dry air mass in the column, surface pressure, base-state &amp; perturbation pressure, zonal &amp; meridional wind at 10 km</td>
<td>Barker et al. (2005)</td>
<td></td>
</tr>
</tbody>
</table>
The WRF-based ensemble Kalman filter (EnKF) system used in this dissertation is same as Meng and Zhang (2008a; b) and Zhang et al. (2009). Details about the EnKF architecture are provided in Table 3.3. The EnKF was run using 50 members whose initial and boundary conditions were perturbed from ERA-interim reanalysis using a domain-specific background error covariance and the “cv5” option of the WRF 3DVar package (Barker et al., 2004). The perturbed variables in the ensemble included perturbation potential temperature, zonal and meridional wind, cloud water mixing ratio, vapor mixing ratio, rain mixing ratio, base-state and perturbation geopotential, base-state and perturbation dry air mass in the column, surface pressure, base-state and perturbation pressure, and zonal and meridional wind at 10 km. Radial velocity was assimilated into both domains every 15 min after WRF spun up for 6 h. The background error covariance was inflated prior to each assimilation using the covariance relaxation method proposed by Zhang et al. (2004, their Eq. 4) with a relaxation coefficient of 0.8. The successive covariance localization (SCL) method (Zhang et al., 2009) was used with a fifth-order correlation function (Gaspari and Cohn, 1999) and a horizontal radius of influence of 90 and 30 km for the outer and inner domains, respectively. The vertical radius of influence was set to six model levels. The EnKF updated perturbation potential temperature, zonal and meridional wind, cloud water mixing ratio, vapor mixing ratio, rain mixing ratio, perturbation geopotential, perturbation dry air mass in the column, surface pressure, perturbation pressure, and zonal and meridional wind at 10 km.

Two sets of simulations were conducted. One of a squall line on 23 December 2011 and one of a non-squall MCS on 16 October 2011. Details about these storms are provided in Section 3.3. Each set of simulations contained three model runs whose only difference was their microphysical parameterization. The same ensemble was used to initialize all simulations on 23 December and 16 October, respectively.
3.2.4 Model Spatial Compositing Technique

Assimilating radial velocity results in each ensemble member developing a mesoscale midlevel inflow. However, details about the midlevel inflow, including its magnitude, slope, length, and location slightly vary among members. This variability is expected and required since the EnKF is a best-linear estimator and assumes that the ensemble represents the range of all possible outcomes. This variability is problematic when the spatial pattern of microphysical processes around the midlevel inflow is analyzed because it smears the spatial pattern and makes its association with the midlevel inflow unclear. Thus, prior to analyzing the spatial pattern of microphysical processes, this dissertation determined if a robust midlevel inflow existed in a given solution and spatially composited all robust midlevel inflows so they were the same horizontal size.

The first step in creating the spatial composites was to identify robust midlevel inflows. At a given time, a set of vertical cross sections was taken from each ensemble member. Cross sections from the 23 December simulations were examined at 1930 UTC, after radial velocity data had been assimilated seven times. Cross sections from the 16 October simulations were analyzed at 1445 UTC, after radial velocity data had been assimilated eleven times. Analysis was done after a different number of assimilation periods in the two sets of simulations since the squall line on 23 December rapidly formed a midlevel inflow but the MCS on 16 October did not form a midlevel inflow until late in its lifecycle (not shown). The analysis has also been conducted at different time steps and by compositing the same cross section at multiple times. In all cases the composites are fundamentally the same (not shown). Thus, the results presented in this chapter are robust. 1930 UTC 23 December and 1445 UTC 16 October were specifically selected since they provided the
largest sample size of cross sections containing robust midlevel inflows. In each of the subsequent figures the number of cross sections included in the composites is listed on the right-hand side of each panel.

The midlevel inflow observed by S-PolKa on 23 December descended predominantly from west to east. The midlevel inflow observed by S-PolKa on 16 October was primarily oriented from south to north. Based on this observed geometry, the cross sections from the WRF simulations were oriented zonally on 23 December and meridionally on 16 October. Given that the midlevel inflow region was broader during the 16 October non-squall MCS, six cross sections were taken from each ensemble member during the 16 October simulations and five cross sections were taken from each ensemble member during the 23 December simulations. Multiple cross sections were considered from each member in order to increase the sample size within the composites. However, each cross section was evaluated independently. Thus, the composites may contain no cross section, one cross section, or multiple cross sections from a given ensemble member. The number of cross sections from each ensemble member does not change the results of this dissertation.

The process of identifying robust midlevel inflows in each cross section and determining the core of the robust inflows was as follows:

1. *Convert the vertical coordinate in the cross sections from their native unevenly spaced eta coordinates to evenly spaced height coordinates.*

2. *Isolate deep convection.* Any vertical column within the cross section that lacked reflectivity greater than 5 dBZ above 5 km was ignored.

3. *Remove convective cores.* In the 23 December simulations, data within vertical columns that had reflectivity greater than 30 dBZ above 8 km and vertical velocities greater than 2 m s\(^{-1}\) at any altitude were ignored. In the 16 October simulations, data within vertical
columns that had reflectivity greater than 30 dBZ above 6 km and vertical velocities greater than 2 m s\(^{-1}\) at any altitude were ignored. The threshold changed due to convective intensity differences in the storms.

4. *Identify the location of the potential midlevel inflow.* Isolate locations within the cross section that had horizontal wind speeds greater than 18 m s\(^{-1}\) during the 23 December simulations or greater than 9 m s\(^{-1}\) during the 16 October simulations. This region was called the potential midlevel inflow. The threshold changed due to wind speed differences in the storms. If no region within the cross section satisfied this criteria the cross section was excluded from the analysis.

5. *At each model level, identify the location of the maximum wind speed within the potential midlevel inflow.* These locations were referred to as maximum points.

6. *Ensure that the potential midlevel inflow was linear.* Use the maximum points to calculate the best fit linear line. Exclude the cross section if the correlation between the maximum points and the best fit line was less than 0.9. It is appropriate to use a best fit linear line since the vertical coordinate of the cross section was converted from unevenly spaced eta values to uniform height intervals.

7. *Ensure that the midlevel inflow was not sloped too steeply.* Exclude the cross section if the slope of the best fit linear line was less than -15 or greater than 0. This ensured that the potential midlevel inflow gradually descended from left to right and ensured that processes associated with large-scale convective updrafts did not contaminate the data used in this chapter. It is appropriate to calculate and analyze the slope of the best fit linear line since the vertical coordinate of the cross section was converted from unevenly spaced eta values to uniform height intervals.
8. Find the base of the core of the midlevel inflow. Find the maximum point that had the lowest altitude. This was called the convective end of the midlevel inflow core. Ignore all data to the right of this point.

9. Ensure that the core of the midlevel inflow was coherent. If a maximum point was more than 0.1° longitude away from the maximum point immediately to its right, ignore that maximum point and all subsequent maximum points to its left. This length scale is arbitrary defined. Results are insensitive to the length scale. Changing the length scale only changes the number of cross sections in the composites.

10. Find the top of the core of the midlevel inflow. Find the maximum point that had the highest altitude and the left most horizontal position. This was called the anvil end of the midlevel inflow core. Any maximum point that was above and to the right of this point was ignored.

11. Ensure that the core of the midlevel inflow was large enough. Exclude the cross section if the distance between the convective and anvil end of the midlevel inflow core was less than 0.2° longitude. This length scale is arbitrary defined. Results are insensitive to this length scale. Changing this length scale only changes the number of cross sections in the composites.

12. Ensure that only one coherent storm was contained within the midlevel inflow core. Exclude the cross section if more than five adjacent model grid points had reflectivity less than 5 dBZ at any vertical level or if there were vertical holes where reflectivity was less than 5 dBZ.
If these criteria were satisfied the cross section was said to contain a robust midlevel inflow and the midlevel inflow core is defined as the region between the maximum points at convective and the anvil end. All subsequent analysis only used data within the midlevel inflow core.

Once all midlevel inflow cores were located, they were arbitrarily scaled to the same length and shifted to the same location. Results are insensitive to the assumed length scale and shifted location. No vertical scaling was conducted since the height profile was nearly the same in each simulation.

One of the primary objectives of this chapter is to compare the spatial pattern of ice processes within the midlevel inflow region from WRF with PID observations. The PID only indicates what process is dominant at a given location. No information about the microphysical production rate is provided. Thus, the PID can only validate the location of microphysical processes. To match the PID data, the spatial pattern of ice microphysical processes generated by the three microphysical parameterizations used in this dissertation was represented in terms of their composite frequency. At each grid point the number of midlevel cores that had a nonzero microphysical process rate was normalized by the total number of midlevel cores. While the PID gives an indication of the dominant microphysical processes at each location, the WRF composites were not generated using the dominant simulated microphysical process since the definition of the dominant process differs in the PID and WRF. In the PID, the dominant microphysical process is the process that produces the hydrometeor that is the largest, densest, or has the highest dielectric constant since these particle accounts for the largest radar return power. The dominant microphysical process in WRF is defined based on the production rate of the process, which cannot be observed using the PID. Given these differences, it is more appropriate to compare the PID to the full simulated microphysical process fields rather than simulated fields of the dominant microphysical process.
In order to ensure that composites were statistically robust, the frequency was only reported if more than half of the cores at that grid point had reflectivity greater than 5 dBZ. Thus, the frequency composites shown in this chapter depict where processes were statistically more and less likely to occur relative to the midlevel inflow.

This chapter also produced composites of radar reflectivity, horizontal wind speed, vertical velocity, and temperature. Similar to above, only grid points where more than half of the cores had reflectivity greater than 5 dBZ were included. The reflectivity used in this chapter was directly output by WRF and was produced by an S-band radar simulator that had been adapted to fit the assumed hydrometeor size distributions in each scheme. All scattering was assumed to be in the Rayleigh regime. WRF simulated reflectivity is directly comparable to observed S-PolKa reflectivity since the S-PolKa radar is also an S-band radar.

3.3 Dual-Polarimetric Observations of Mesoscale Convective Systems

Chapter 2 identified 36 MCSs during DYNAMO/AMIE that contained a midlevel inflow and created spatial composites that showed that the S-PolKa PID data had a systematic spatial pattern relative to the midlevel inflow. Squall line and non-squall MCSs were composited separately in Chapter 2 and the organization of the PID data relative to the midlevel inflow was unchanged by whether or not the MCS was organized into a squall line. This chapter simulated a squall line and non-squall MCSs that were included in the composites created in Chapter 2. It is important that this chapter considers two different types of MCSs in order to investigate if the morphological structure impacts the simulated spatial pattern of ice microphysical processes. The squall line
simulated in this chapter occurred on 23 December 2011 (Figure 3.2a-d). This case was selected since it was one of the most well-developed and strongest squall lines observed by S-PolKa during DYNAMO/AMIE. At 1935 UTC a broken convective line was oriented north-south and its leading edge was about 100 km to the east of the S-PolKa radar (Figure 3.2a). Stratiform precipitation extended from the convective line westward for 100-200 km. Figures 3.2b-d show a vertical cross section of reflectivity, radial velocity, and PID data along the red line in Figure 3.2a. Reflectivity shows that this storm had a well-defined convective zone characterized by vertically erect convective cores, a transition region characterized by weaker reflectivity, and a stratiform region with a distinct brightband. Radial velocity shows that the squall line had a strong descending midlevel inflow whose radial velocities exceeded 20 m s\(^{-1}\) approximately 50-75 km away from the S-PolKa radar. This midlevel inflow extended from beneath the anvil (approximately 15 km from S-PolKa) into the stratiform region and provided the primary forcing for the squall line. PID data indicate that the frozen hydrometeors had a layered structure with small ice crystals at cloud top, dry aggregates at midlevels above the melting level, and wet aggregates near the melting level. Additionally, graupel/rimed aggregates were located in few isolated regions along the upper boundary of the wet aggregate layer. The isolated nature of the graupel/rimed aggregates is not a feature unique to this case. Whenever graupel/rimed aggregates was observed within one of the 36 midlevel inflows regions investigated in Chapter 2 these hydrometeors always occurred in shallow, isolated pockets. The cross section shown in Figures 2b-d was selected since it clearly shows the midlevel inflow and layered microphysical pattern within the stratiform region. However, these features were not unique to this cross section. Any cross section taken within this portion of the 23 December squall line contained these characteristics. Thus, the overall hydrometeor pattern observed during the 23 December squall line is consistent with the systematic organization
Figure 3.2: (a) Horizontal map of column maximum S-PolKa reflectivity at 1935 UTC on 23 December 2011. (b) Vertical cross section of S-PolKa reflectivity taken along the red line in (a). (c) Vertical cross section of S-PolKa radial velocity taken along the red line in (a). (d) Vertical cross section of S-PolKa PID data taken along the red line in (a). (e–h) Same as (a–d) except at 1450 UTC on 16 October 2011.
identified in Chapter 2 and long thought to exist in oceanic tropical convective systems (Leary and Houze, 1979b; Houze, 1989). Interpreting the PID in terms of microphysical processes suggests that ice processes during this storm were layered with only depositional growth near echo top, aggregation at midlevels above the 0°C level, and melting and shallow pockets of riming of near the 0°C level. At the levels where aggregation was occurring, the larger particles dominated the radar signal, thus preventing deposition from being identified. However, because upward motion was likely occurring above the midlevel inflow layer, the aggregates would have been also accumulating mass by depositional growth. Therefore, the layer where aggregation is occurring is interpreted as a zone where both aggregation and vapor deposition were active. This interpretation is important when comparing the PID with model output since the parametrizations calculate aggregation and deposition separately and therefore keep track of both processes. A similar interpretation applies to riming. If rimed particles are present in a zone of upward motion, they too would be growing by vapor deposition as well as riming. The cross section shown in Figures 2b-d was selected since it clearly showed the midlevel inflow and layered microphysical pattern within the stratiform region. However, these features were not unique to this cross section. Any cross section taken within this portion of the 23 December squall line contained these characteristics.

The non-squall MCS simulated in this chapter persisted over the S-PolKa radar for nearly 18 h on 16 October 2011. This MCS was selected since it was one the largest MCSs observed by the S-PolKa radar during DYNAMO/AMIE. Figures 3.2e-h shows the storm at 1450 UTC as a midlevel inflow developed during the later stages of this MCS. Prior to this time this MCS lacked a midlevel inflow. Thus, our analysis could not be conducted when the 16 October MCS was in its developing or mature stages. The convective and stratiform regions were organized less
systematically with Figure 3.2e showing pockets of high reflectivity convective cores scattered within an expansive region of weaker reflectivity stratiform precipitation. Figures 3.2f-3.2h show a cross section of reflectivity, radial velocity, and PID data through a portion of the stratiform precipitation along the red line in Figure 3.2e. This particular cross section was selected since it clearly showed the midlevel inflow and systematic layered hydrometeor/microphysical pattern that was commonly observed in the stratiform region of MCSs. However, these patterns were not unique to this cross section, any cross section taken within the southeast portion of 16 October MCS around 1450 UTC displayed these characteristics.

While some hydrometeor/microphysical differences exist between the 23 December squall line and 16 October non-squall MCS, the S-PolKa PID data had the same fundamental spatial pattern relative to the midlevel inflow in both cases. The 16 October non-squall MCS was shallower than the 23 December squall line (Figures 3.2b, 3.2f) and its precipitation was weaker. A weak brightband was apparent on 16 October, but a portion of the precipitation did not reach the surface. Similar to 23 December, 16 October had a well-defined midlevel inflow (Figure 3.2g). However, the midlevel inflow on 16 October remained at a constant altitude of 5 km and only had a maximum intensity of 15 m s\(^{-1}\). Loops of reflectivity and radial velocity (not shown) suggest that the midlevel inflow on 16 October accelerated the dissipation of the MCS, instead of forcing the MCS as observed on 23 December. A few differences in the detailed spatial patterns of the ice processes existed between these storms. Unlike the 23 December squall line, PID data during the later stages of the 16 October MCS often lacked shallow pockets of graupel/rimed aggregates and the dry aggregate layer was much shallower. This comparison suggests that the stratiform region associated with the 16 October non-squall MCS lacked riming and had less aggregation than the stratiform region associated with the 23 December squall line, which may be attributable to weaker
vertical motions on 16 October. However, these riming and aggregation differences were relatively minor, and both cases were dominated by a layered pattern consistent with Chapter 2. Thus, both MCSs were characterized with a layered ice process pattern that transitioned from mostly deposition in upper levels to aggregation to melting with increasing distance from echo top.

This dissertation compares these two snapshots of PID data from 23 December and 16 October with composites of simulated microphysical patterns from three microphysical parameterizations in WRF. It is appropriate to compare WRF composites to these PID examples since it has been shown that these PID snapshots are representative of the statistically robust microphysical pattern detailed in composites created in Chapter 2.

### 3.4 Kinematic Structure of Simulated Mesoscale Convective Systems

Before the spatial pattern of simulated ice processes can be compared with the PID, it must be proven that assimilation successfully results in the ensembles developing a mesoscale circulation consistent with observations. Figure 3.3 shows cross sections of S-PolKa radial velocity and composite simulated horizontal wind speed for the 23 December and 16 October MCSs in shading. S-PolKa radial velocity and WRF horizontal wind speeds are not directly comparable since radial velocity describes a portion of the horizontal and vertical wind component. However, within the midlevel inflow region the vertical velocity component is often an order of magnitude smaller than the horizontal velocity component. Streamlines calculated within the plane of the WRF cross sections are oriented nearly horizontal (not shown). While S-PolKa radial velocity underestimates horizontal winds, this underestimation is minimized since cross sections of S-PolKa data were
oriented along the dominant direction of the midlevel inflow (see section 2.4). Additionally, the underestimation by S-PolKa does not impede these results since validating the absolute magnitude of the simulated midlevel inflow is not an objective of this dissertation. The only requirement is that the simulated midlevel inflows to have a magnitude similar to observations. Thus, it is appropriate to compare the WRF horizontal wind speeds to S-PolKa radial velocity. Figure 3.3 shows that each ensemble had a composite midlevel inflow whose slope and magnitude was similar to the S-PolKa observations.

Figure 3.3: (a) Cross section of the average S-PolKa radial velocity at 1930 UTC on 23 December 2011. (b) Cross section of the composite average horizontal wind speeds simulated by the Milbrandt-Yau (MY) scheme at 1930 UTC on 23 December 2011 in shading. The composite average reflectivity is shown in gold contours with reflectivity values contoured every 5 dBZ from 5 to 45 dBZ. The boundaries of the midlevel inflow core are marked by black vertical lines. The position of the midlevel inflow is approximated by the dashed diagonal magenta line. The magenta number along the right side indicates how many cores are included in these composites. (c) Same as (b) except for the Morrison (MOR) scheme. (d) Same as (b) except for the WDM6 (WD) scheme. (e–h) Same as (a–d) except taken at 1445 UTC on 16 October 2011.
The shading in Figure 3.4 shows the composite simulated vertical velocities on 23 December and 16 October. The midlevel inflow region is characterized by a broad weak updraft above the midlevel inflow and a broad weak downdraft below the midlevel inflow, consistent with previous observations (Houze 2004; 2014). Note that the color bars in Figure 3.4 have different scales. The scale used for 23 December is twice as large as the scale used for 16 October. The 23 December ensembles were expected to have consistently greater vertical velocities than the 16 October ensembles since the 23 December case was an active squall line and the 16 October case was a late stage MCS. However, the magnitude of these vertical velocities cannot be validated since in situ vertical velocity data was unavailable during these storms.

While each simulation is characterized by large-scale ascent above the midlevel inflow and descend below the midlevel inflow, small-scale vertical velocity differences in Figure 3.4 indicates that convective-scale differences in the simulated vertical velocity pattern exist despite data assimilation. Given that the airflow through these storms was dominated by the midlevel inflow, this convective-scale vertical velocity variability is not a serious problem. Thus, the comparison of the observed and simulated spatial pattern of ice processes within the midlevel inflow region proceeds with confidence that most of the microphysical differences exist due to differences in the processes and not because of large-scale dynamical differences. The impact of convective-scale vertical velocity variability on the results will be discussed below.
Figure 3.4: (a) Cross section of the composite average vertical velocity simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average reflectivity is shown in gold contours. The contour interval is every 5 dBZ from 5 to 45 dBZ. The edges of the core of the midlevel inflow are marked by black vertical lines. The composite average 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top green lines, respectively. The position of the midlevel inflow is approximated by the dashed diagonal magenta line. The magenta number along the right side indicates how many cores are included in these composites. (b) Same as (a) except for the Morrison (MOR) scheme. (c) Same as (a) except for the WDM6 (WD) scheme. (d –f) same as (a –c) except at 1445 UTC on 16 October 2011.
3.5 Model Comparison

3.5.1 Deposition

Figure 3.5 shows the frequency of deposition within the midlevel inflow regions on 23 December and 16 October in shading. In both sets of simulations, each of the three microphysical parameterizations indicated that deposition was possible anywhere above an altitude of 5 km, which is approximately the 0°C level. This spatial pattern is consistent with Figure 3.2 and composites presented in Chapter 2, which show a layer of dry aggregates and graupel/rimed aggregates topped by a layer of small ice crystals, all of which are inferred to be growing by deposition in an environment of in-cloud upward air motion. Additionally, this simulated pattern is consistent with laboratory and field observations that show that ice can grow via deposition between 0° and -70°C (e.g. Wallace and Hobbs, 2006; Bailey and Hallett, 2009) and has been observed in previous modeling studies including Caniaux et al. (1994) and Donner et al. (2001).

The spatial pattern of simulated deposition varied in both sets of simulations. Deposition did not occur everywhere above the 0°C level, each parameterization had regions statistically more and less likely to have deposition. Additionally, the location of the maximum and minimum frequency of deposition differed among the parameterizations. This spatial variability was linked to variations in the vertical air motion. Deposition was always collocated with regions of upward vertical motion (yellow shading and orange lines in Figure 3.5). Because upward motion was statistically more frequent above the midlevel inflow and near echo top,
deposition was more common near the echo top. The frequency of deposition increased the least near echo top in MOR on 23 December, because the upward vertical velocity was statistically weaker (Figure 3.4). The relationship between deposition and upward motion agrees with previous studies concluding that ice particles grow via deposition within the mesoscale updraft zone (e.g.

Overall, deposition occurred more frequently in each of the three parameterizations on 16 October than 23 December despite vertical velocities being weaker on 16 October than 23 December. This difference suggests that vertical motion on 16 October was more uniform and had less small-scale variability, which is expected since the 16 October MCS was in a later stage of its lifecycle than the 23 December MCS (Figures 3.3d, 3.3h, and 3.4).

3.5.2 Aggregation

The composite frequency of aggregation is shown on 23 December and 16 October in shading in Figure 3.6. Aggregation had the most consistent spatial pattern of the microphysical processes analyzed in this chapter with all parameterizations producing aggregation everywhere above the 0°C level. While MORE had a slightly reduced frequency of aggregation just above the 0°C level, the frequency of aggregation in MY and WD was almost uniformly one everywhere above the 0°C level. There was no significant difference in the frequency of simulated aggregation between the 23 December and 16 October MCSs.

While the microphysical parameterizations were consistent with the PID in indicating that aggregation occurred near the 0°C level, serious discrepancies existed in terms of the maximum depth of the aggregation layer. Each simulation has aggregation always occurring at temperatures below -20°C and reaching echo top and. Figure 3.2 indicates that aggregation never extended to echo top on 23 December or 16 October, a pattern that was also observed in each of the 36 midlevel
inflows analyzed in Chapter 2. Additionally, this simulated pattern disagrees with laboratory and observational studies that demonstrated that aggregation rarely occurs below -20°C (e.g. Hobbs et al., 1974). Studies such as Houze and Churchill (1987), and Braun and Houze (1994) indicate that aggregation is most common in the stratiform portion of an MCS within 1-2 km of the melting level.
One factor in this altitude discrepancy between the simulations and PID may be related to the fact that simulated aggregation is defined in this chapter to be any process in which frozen particles collect each other. This includes processes in which very small ice crystals collide and coalesce, which is likely undetectable by the PID. However, based on laboratory studies even this weak aggregation is not expected at very low temperatures (e.g. Hobbs et al., 1974). Thus, a systematic error in the representation of aggregation at low temperatures likely exists in these three parameterizations. This abundance of aggregation may have serious implications for simulated convection by impacting radiative fluxes and causing ice particles to become too large and fall too quickly. Van Weverberg et al. [2013] found that simulated tropical MCSs were very sensitive to the fall speed of frozen hydrometeors and accurate simulations required small, slowing falling particles. Roh and Satoh [2014] demonstrated that shallow and midlevel convection were reduced and deep MCSs with stratiform precipitation were increased in simulations using a single-moment microphysical parameterization by preventing snow and ice from accreting onto graupel. These accretion variables are included within the aggregation term used in this dissertation. Thus, the parameterization of aggregation in these schemes may be a factor in limiting the ability for numerical simulations to accurately simulate stratiform precipitation. Modeling studies including Morrison et al. (2009), Hagos et al. (2014), and others extending back to Fovell and Ogura (1988) have documented the deficiency of stratiform precipitation in simulations. Evidence of deficient stratiform precipitation in these simulations will be presented in section 3.6.
3.5.3 Riming

Figure 3.7 shows the frequency of riming on 23 December and 16 October in shading. Each parameterization indicated that both cases were characterized by frequent riming near the 0°C level, and that riming occasionally extended to altitudes as great as 10 km. It is important to keep in mind that simulated riming in this dissertation includes processes that create water-coated frozen hydrometeors, which may be partially responsible for the high frequency of riming near an altitude of 5 km in the simulations (i.e. the approximate location of the 0°C level). Despite these broad similarities, riming was the process analyzed in this chapter that displayed some of the largest

Figure 3.7: Same as Figure 3.5 except for simulated riming.
differences among individual parameterizations and between the two MCS simulations. A striking difference between the schemes in both cases was the presence of riming in MY well below an altitude of 5 km, which is physically impossible since the 0°C level is at 5 km. Frozen hydrometeors were never observed near the surface during DYNAMO/AMIE, and indeed are never observed over oceanic environments near the equator. While definition of riming in this chapter includes processes that could result in water-coated frozen hydrometeors, the frozen core of these hydrometeors are still expected to melt rapidly below the 0°C level and should not be present well below the 0°C level. Thus, the near surface riming in MY is clearly erroneous. However, while Figures 3.7a and 3.7d suggest MY had frequent riming below the freezing level, the rate of riming was very low (not shown), so this behavior is a model artifact that does not substantially affect the results of this dissertation or MY simulations.

Riming in each scheme on 23 December extended to greater altitudes near the convective end than the anvil end of the midlevel inflow (left compared to right side of each panel top row Figure 3.7). This spatial riming pattern is consistent with simulations conducted by Caniaux et al. (1994). The schemes differed in how fast riming reduced with height above the melting level and the amount of riming near the 0°C level in the anvil end of the midlevel inflow. Peak riming in MY and WD was deeper near the convective end and extended as a narrow band through the entire anvil end. The narrow band produced by each of these simulations may be impacted by the inclusion of processes that create water-coated frozen hydrometeors in our definition of riming. Peak riming in MOR was shallower and concentrated in the convective half of the midlevel inflow; riming was rare in the anvil half.

Riming occurred within a shallower layer and had a more consistent spatial pattern among the parameterizations on 16 October (bottom row of Figure 3.7). This difference between the cases is
expected since the 16 October non-squall MCS was in a later stage of its lifecycle and had less intense, more uniform vertical motions than the 23 December active squall line (Figures 3.2d, 3.2h, and 3.4). Each parametrization on 16 October was characterized by frequent rimening over the length of the midlevel inflow near 0°C level, a pattern which may be partially produced by the inclusion of processes that create water-coated frozen hydrometeors in the definition of rimening in this chapter. The frequency of rimening decreased with height in each scheme. This decrease was most rapid in the WD.

Parameterized rimening was not strongly correlated with the mesoscale vertical velocity in this dissertation (orange contours in Figure 3.7). Rimening requires supercooled water, which is produced through stronger updrafts or turbulence. Chapter 2 suggested that the rimening within the midlevel inflow region during DYNAMO/AMIE could have resulted from shear-induced turbulence or small embedded convective cells or may have been residual of a more convective phase of the storm. In any case, convective-scale vertical motions are the most likely source of supercooled water in these simulations. This convective nature would make the pockets of supercooled water and associated rimening intermittent on scales smaller than the mesoscale midlevel inflow current around which the results were composited around. Given that convective-scale vertical motion varied in these simulations, discrepancies among the schemes in the rimening pattern would not be unexpected.

Similar to the DYNAMO/AMIE PID data and previous studies, each parameterization indicated that rimening occurred near the 0°C level within the midlevel inflow. The presence of rimened particles in the stratiform region of MCSs was first discussed by Leary and Houze (1979b). They inferred the presence of rimening in this layer indirectly from drop-size distributions measured below the brightband. Since then observational and modeling studies have identified rimened particles
within the stratiform precipitation in the Indian Ocean (Chapter 2; Suzuki et al., 2006; Rowe and Houze, 2014), equatorial maritime continent (Takahashi and Kuhara, 1992; Takahashi et al., 1995), West Africa (Evaristo et al., 2010; Bouniol et al., 2010), Oklahoma (Zrnić et al., 1993), Taiwan (Jung et al., 2012), and Europe (Hogan et al., 2002). It is a positive result that all three microphysical schemes identify this riming layer.

However, there are discrepancies in the detailed spatial pattern of riming that are important. The most systematic and significant discrepancy between the parameterizations and PID was the frequency of occurrence of riming and in the depth of the layer in which riming was taking place. In each parameterization, riming extended over a layer that was too deep and/or occurred too frequently. Nearly 70% of the cores in each scheme on 23 December had riming reach an altitude of 8 km somewhere. Figure 3.2d indicates that riming was detected only in very shallow pockets within the midlevel inflow and reached just a few kilometers above the 0°C level in the convective core of the squall line (approximately 115 km from S-PolKa in Figure 3.2d). These shallow pockets of riming are not a feature that is unique to the 23 December squall line. Chapter 2 and Rowe and Houze (2014) found that all instances of graupel/rimed aggregates within the stratiform portion of MCSs were concentrated near the 0°C level and their vertical extent was limited. Leary and Houze (1979b) inferred that the riming should only be present in this thin layer. In general, riming is expected to be shallow in these storms since they occurred over the near equatorial ocean where vertical motions are weaker (e.g. LeMone and Zipser, 1980; Zipser and LeMone, 1980; Mohr and Zipser, 1996; Nesbitt et al., 2000). This excessive riming may create too many rimed hydrometeors and cause too much latent heat release, which may contribute to previous studies showing a tendency for simulated oceanic convection to be too strong and have too much graupel (e.g.
Wiedner et al., 2004; McFarquhar et al., 2006; Blossey et al., 2007; Lang et al., 2007; Li et al., 2008; Lang et al., 2011; Hagos et al., 2014; Varble et al., 2014).

Figures 3.7d-3.7f indicates that each parameterization produced riming somewhere along the length of the midlevel inflow in at least 70% of the data analyzed on 16 October. PID observations on 16 October consistently indicated that riming was rare in its stratiform region. While the PID may miss regions of weak riming, the vertical motions and turbulence were likely weak at this late stage of the storm's lifetime and would therefore not be expected to support frequent riming. Thus, the microphysical parameterizations appear to be creating too much riming near the 0°C level.

3.5.4 Melting

The frequency of melting on 23 December and 16 October is shown in Figure 3.8. In each set of simulations, these parameterizations indicated that melting occurred near the 0°C (bottom orange line in Figure 3.8). The depth of melting was smallest in MOR with melting frequently isolated to one model level. Melting occurred over a somewhat deeper layer in WD. The spatial pattern of melting in MY was distinctly different. Melting in MY almost always reached the surface on 23 December and often extended down to an altitude of 2 km on 16 October. Again, MY allowed ice well below the melting layer, which is a behavior never actually observed in the oceanic tropics. However, the near-surface melting in MY occurred at a very low rates (not shown), again indicating that it is a model artifact that does not affect the basic conclusions of this dissertation.
Excluding MY, the simulated spatial pattern of melting was the most consistent with PID observations and previous studies of any process analyzed in this chapter. PID data close to the S-PolKa radar, where the impact of beam broadening was minimal, indicated that melting occurred in a very narrow band. Previous studies have suggested that melting in the stratiform region occurs within 1 km of the 0°C level (e.g., Leary and Houze, 1979b; Houze, 1981; Tao and Simpson, 1989; Caniaux et al., 1994; Donner et al., 2001). Thus, while MOR and WD were both consistent with the PID and previous studies, the shallower layer of peak melting frequency in MOR was slightly more accurate than the deeper layer in the WD simulation.
3.6 Impact of Microphysical Differences

From the foregoing discussions in this chapter, it is evident that no single parameterization accurately portrays the precise spatial distribution of microphysical processes in either of the two MCSs considered. All the schemes reproduce the right basic layering of microphysical processes, but disparities in the detailed spatial patterns are prevalent. Do these spatial dissimilarities imply substantial differences in the reflectivity (i.e. precipitation structure) between the parameterizations?

The right three columns of Figure 3.9 shows the composite reflectivity cross section for 23 December and 16 October. For comparison, a composite S-PolKa reflectivity cross section is shown in the left column. Figure 3.10 shows the corresponding horizontal maps of reflectivity at an altitude of 5 km. While each parameterization correctly simulated a squall line with a convective and stratiform end on 23 December and an expansive region of weak precipitation on 16 October, each parameterization exhibited different patterns of reflectivity and none matched S-PolKa observations.

MY and WD were similar on 23 December with both having a well-defined anvil in Figures 3.9b and 3.9d. However, the reflectivity near the convective end of the midlevel inflow was stronger in MY (Figure 3.9b) and the brightband was more distinct in WD (Figure 3.9d). Figures 3.10b and 3.10d show that WD produced a better-defined leading convective line on 23 December than MY. However, these figures also demonstrate that both simulations produced too little stratiform precipitation, a trend that has been consistently reported since Fovell and Ogura (1988). MOR was most similar to S-PolKa observations on 23 December. It had the most distinct
convective, transition, and stratiform regions (Figure 3.9c) and the horizontal distribution of reflectivity was most accurate (Figure 3.10c).

The parameterizations had even more difficulty representing the reflectivity structure on 16 October. As in the 23 December simulations, the MOR simulation was most similar to S-PolKa observations, with a well-defined bright band (Figure 3.9g) and isolated convective pockets whose locations were similar to observations (Figure 3.10g). However, the reflectivity in MOR on 16 October was too high near the surface in the convective end of the composites (left side of Figure 3.9g) and across the entire anvil end of the composites (right side of Figure 3.9g). Additionally, reflectivities within the isolated convective pockets on 16 October tended to be too high in MOR (Figure 3.10f). MY lacked a brightband on 16 October and the reflectivity on the anvil end the
composites was often too high (Figure 3.9f). At the 5 km level, reflectivities simulated by MY were often too low (Figures 3.10g), reflecting the fact that the brightband in MY was too weak. While the horizontal distribution of reflectivity from WD was similar to S-PolKa on 16 October (Figure 3.10h), its reflectivity cross section drastically differed from S-PolKa and the other schemes. Figure 3.9h shows that WD had a strong brightband across the entire midlevel inflow and reflectivity was not observed below 3 km on 16 October. Thus, WD on 16 October was extremely deficient in stratiform precipitation. This result is consistent with Varble et al. [2011] and Hagos et al. [2014], who found that microphysical parameterizations in WRF tended to
accurately simulate the areal coverage of stratiform precipitation in tropical MCSs but underestimated their stratiform rainfall rates. The increased discrepancy among the parameterizations and observations on 16 October as compared to 23 December is not surprising since the forcing on 16 October was weaker and the MCS was characterized by more stratiform precipitation. Studies such as Morrison et al. (2009) and Hagos et al. (2014), and others extending back to Fovell and Ogura (1988), have consistently demonstrated that models have difficulty simulating the stratiform precipitation of MCSs.

As a unit Figures 3.9 and 3.10 indicate that large reflectivity discrepancies were witnessed among the parameterizations even though data assimilation forced each ensemble to have the same mesoscale circulation and the spatial distributions of microphysical processes were similar to first order. It should be noted that the spatial pattern of the ice-phase processes were not the only differences among the parameterizations. The schemes were characterized by different assumed drop size distributions and fall speeds. Additionally, MY was the only scheme that simulated hail in addition to low-density graupel. (Although, it is important to note that hail in MY only occurred in a few ensemble members and was extremely isolated.) While these parameter differences contribute to variations in the simulated spatial distribution of precipitation in both the horizontal and vertical dimensions, the details of the spatial distributions of the ice-phase microphysical process must also factor into obtaining the correct precipitation distribution. The detailed differences in both the liquid and ice microphysics between parameterization schemes result in different interactions with the convective-scale dynamics, which in turn affect the mesoscale evolution of the storms. The differences in dynamical/microphysical evolution manifest themselves in the magnitude and location of heating and cooling processes in the storms (Tao and Simpson, 1989). Szeto et al. (1988) and Tao et al. (1995) have suggested that squall line
simulations are especially sensitive to simulated melting patterns. Van Weverberg et al. (2014) systematically changed MOR so that its parameterization of microphysical processes exactly matched MY. In the process they showed that differences in the parameterization of microphysical processes accounted for structural differences between the schemes. Thus, while the spatial pattern of ice processes is not the only reason why convection in WRF varies among parameterizations, it is likely an important factor. Simulations will likely become more accurate as the spatial representation of these processes are improved (Grasso et al., 2014).

### 3.7 Conclusions

Chapter 2 used the NCAR particle identification (PID) algorithm to demonstrate that ice hydrometeors in the midlevel inflow region of mesoscale convective systems (MCSs) during DYNAMO/AMIE had a systematic spatial pattern with sequential layers of small ice crystals, dry aggregates, and wet aggregates descending from echo top and shallow pockets of graupel/rimed aggregates just above the melting level. This overall pattern was observed in all types of MCSs. While PIDs have been traditionally interpreted as an indication of the dominant hydrometeor, the frozen categories identified by the PID can be interpreted as ice-phase microphysical processes. In particular, the category identified as small ice crystals represents populations of particles growing mostly by vapor deposition. Dry aggregates represent zones with particles that are redistributing their mass through aggregation while still increasing in mass via vapor deposition. The category referred to as graupel/rimed aggregates denotes regions where the ice particles are predominantly increasing in mass as supercooled water freezes to their surfaces in a process referred to as riming.
These particles are also likely simultaneously increasing in mass due to deposition. The category called wet aggregates indicates where ice particles are melting. The results from Chapter 2 therefore indicate that these basic ice-phase microphysical process are distributed in the midlevel inflow zone of tropical, oceanic MCSs in a systematic layered pattern.

This chapter investigates whether the spatial pattern of ice process simulated by three routinely available microphysical parameterizations in the Weather Research and Forecasting (WRF) model are consistent with the patterns obtained from radar data in Chapter 2, our theoretical understanding of microphysical processes, and each other. The microphysical parameterizations included were the Milbrandt-Yau Double-Moment scheme (MY), the Morrison 2-Moment scheme (MOR), and the WRF Double-Moment 6-Class scheme (WD). Radial velocity data were assimilated into WRF so that all simulations would develop a mesoscale circulation similar to observations and ensure that microphysical differences were not caused by large-scale dynamical differences. Cross sections through each storm were selected based on the robustness of their midlevel inflow and were spatially composited so that slight variations among the individual cross sections did not smear the spatial microphysical pattern. Microphysical processes were obtained by summing the variables within each scheme that calculated the rate at which the mixing ratio of individual particles interact according to the process involved. A squall line and non-squall MCS were simulated in order to test if the simulated spatial pattern of the ice microphysical processes were dependent on the morphology of the MCS.

The three microphysical parameterizations had a broad spatial ice microphysical pattern that was consistent with the PID, each other, and our theoretical understanding of microphysics. These consistencies included:
• The parameterizations had the same process at a given location within the midlevel inflow as the PID. All were characterized by deposition anywhere above the 0°C level, aggregation at and above 0°C level, melting at and below the 0°C level, and riming near the 0°C level.

• Each parameterization had deposition located above the 0°C level wherever upward motion existed. The PID indicates ice particles (small ice crystals, dry aggregates, or graupel/rimed aggregates) existed everywhere above the 0°C level. These particles would be expected to be growing by vapor deposition since upward air motion was generally occurring above the midlevel inflow layer.

• The small-scale spatial variability of simulated riming is likely caused by differences in the convective-scale vertical velocity motions. Chapter 2 suggested that the riming occurred in pockets within the midlevel inflow region due to small-scale shear induced turbulence or embedded or residual convective cells.

• Except for MY, melting was the microphysical process with the greatest consistency. Melting in the PID, MOR, and WD was concentrated in a narrow band at the 0°C level, which agrees with previous observational and modeling studies (e.g., Leary and Houze, 1979b; Houze, 1981; Tao and Simpson, 1989; Caniaux et al., 1994; Donner et al., 2001).

• The overall spatial pattern of the microphysics did not depend on whether or not the MCS was organized into a squall line.

Details within these general microphysical patterns varied considerably among the parameterizations, differing both from each other and from the PID. The main differences were:

• None of the schemes produced similar details in every microphysical spatial pattern considered in this chapter. While the spatial pattern of deposition and aggregation were
most similar in MY and WD, the spatial pattern of riming and melting were most similar in MOR and WD.

- Aggregation simulated by each parameterization almost always existed everywhere above the 0°C level, regardless of the life stage of the MCS and temperature, whereas the PID indicated that aggregation never occurred near echo top and extended only a short distance above the 0°C level, especially in the later stage MCSs. Simulated aggregation was notably inconsistent with observational studies, which show that aggregation is nearly impossible below -20°C (e.g. Hobbs et al., 1974). Aggregation simulated at these lower temperatures could create biases in the radiative flux and be inimical to the development of stratiform precipitation (e.g. Varble et al., 2011; Van Weeverberg et al., 2013; Hagos et al., 2014; Roh and Satoh, 2014).

- The PID showed riming in the midlevel inflow region was relatively rare, especially during the non-squall late-stage MCS, and only occurred in shallow pockets just above the 0°C level. In general, riming was very common in all simulations, especially near the 0°C level, and sometimes reached altitudes as high as 10 km. Too much riming can be expected to have adverse dynamical feedbacks since too much latent heat is released. However, it is important to note that simulated riming is defined in this dissertation to include processes that create water-coated frozen hydrometeors, which may account for the high riming frequency near the 0°C level in these simulations.

- Rimming and melting in MY were often located well below the 0°C level, which is incorrect; frozen hydrometeors never occurred near the surface over equatorial oceans. However, the degree to which these processes occurred in the model was relatively slight. It is probably an easily correctable problem.
Houze (1989) presented a conceptual model that suggested that ice microphysical processes above the 0°C level within the stratiform portion of an MCS have a layered pattern with deposition aloft, and aggregation combined with deposition and riming between 0 and –12°C. This pattern is generally similar to the PID results in Figures 3.2d and 3.2h and the composites in Chapter 2. Leary and Houze [1979b] concluded that the riming was occurring in a rather shallow layer just above the melting layer. The WRF results presented in this dissertation generally support this layered pattern discussed in these observational and theoretical studies.

While rimed particles have been identified within stratiform precipitation in convective systems in a variety of locations around the globe, the source of these particles has been debated (Chapter 2, Takahashi and Kuhara, 1992; Zrnić et al., 1993; Takahashi et al., 1995; Hogan et al., 2002; Suzuki et al., 2006; Evaristo et al., 2010; Bouniol et al., 2010, Jung et al., 2012; Rowe and Houze, 2014). The simulations presented in this chapter support the theory that convective-scale velocity perturbations are important to the development of riming within the midlevel inflow since the spatial riming pattern varies among the parameterizations despite having similar mesoscale circulation patterns. However, whether the vertical velocity perturbations are shear or buoyancy induced remains to be determined.

This chapter has shown that even while data assimilation requires simulations to evolve with mesoscale circulations similar to those observed, changing the microphysical parameterization creates substantial differences in the details of the horizontal and vertical reflectivity patterns. Given that microphysical processes influence the structure of latent and radiative heating, improving microphysical parameterizations will improve the accuracy of simulated convection at the convective- and global- scale. Until the representation of microphysical processes is improved and made more consistent, simulations will continue to struggle to accurately represent convection.
and results will depend on the parameterization. These new results add to previous findings that demonstrate that the simulated structure of MCSs is sensitive to ice processes (Chen and Cotton, 1988; Szeto et al., 1988; Tao et al., 1991; Tao et al., 1995) and motivate continued ongoing efforts to improve microphysical schemes in cloud resolving models.
CHAPTER 4

DISSERTATION CONCLUSIONS

Radar data have played a crucial role in developing our understanding of mesoscale convective systems (MCSs) and their influence on the global circulation over the last 50 years. MCSs are long duration cloud complexes whose temporal and spatial scale is greater than any of their individual components. While the TIROS satellites in the 1960s provided some of the first data to suggest that tropical oceanic convection is often organized into mesoscale complexes (e.g. Anderson et al., 1966), the structure and quantitative importance of MCSs was not fully appreciated. This changed when quantitative radar reflectivity data were collected aboard four ships during the GARP Atlantic Tropical Experiment (GATE) in 1974. The radar reflectivity data from GATE revealed that the stratiform cloud shield is dynamically active and associated with significant radiative and latent heating fluxes (e.g. Houze, 1977, 1982; Gamache and Houze, 1982). Based on knowledge of the latent heating structure within MCSs obtained during GATE, Hartmann et al. (1984) and Schumacher et al. (2004) were able to demonstrate that the top-heavy heating profile attributed to the stratiform portion of MCSs impacts the global circulation. Barnes and Houze (2013) and Barnes et al. (2015) demonstrated that MCSs are an important component of the net latent heating budget. While Zipser (1969) and modeling studies including Moncrieff (1992) were some of the first studies to suggest that MCSs are associated with a mesoscale circulation, this fact was not
definitively established until Doppler radars were deployed during TOGA COARE (Mapes and Houze, 1995). Using Doppler airborne radar collected in TOGA COARE, Kingsmill and Houze (1999a) showed that air moves through MCSs in distinct three-dimensional layers. Convective cores are characterized by a steeply rising current of air originating in the boundary layer. The stratiform region is characterized by a midlevel inflow that originates at midlevels in the anvil and gradually descends as it flows toward the center of the storm. Houze et al. (2000) and Mechem et al. (2006) showed that the MCSs are important transporters of momentum.

Knowledge of the microphysical processes and their organization within MCSs is important since they link the diabatic and kinematic structures within MCSs and influence the nature of convection at scales ranging from convective to global. Ideas regarding how microphysical processes are organized within tropical oceanic MCSs began to be developed in the late 1970s and 1980s (e.g. Leary and Houze, 1979; Houze 1981; 1989, Churchill and Houze, 1987). While these theories suggest that the organization of microphysical processes within MCSs are related to their mesoscale circulations, the exact nature of the relationship up to now has not been established in observations or numerical simulations. Validating these theories using observations has been difficult since previous sources of microphysical data were limited in their spatial and/or temporal coverage or created biased results (e.g. Comstock et al., 2007; Baumgardner et al., 2011; Korolev et al., 2011). From a modeling perspective, Caniaux et al. (1994) is one of the only studies that indicates how microphysical processes are spatially organized within simulated convection. However, their study used an idealized, two-dimensional model that was unconstrained by observations and did not accurately account for the interaction between the microphysics and dynamics. Their methodology was adopted since it is difficult to sort out dynamical and microphysical causative factors when the microphysics and dynamics are allowed to interact
freely. Hence, how microphysical processes are organized within three-dimensional full-physics numerical simulations has remained largely known.

The Dynamics of the Madden-Julian Oscillation/ARM MJO Initiation Experiment (DYNAMO/AMIE) presented an opportunity to use the latest radar technology to advance our knowledge of MCSs and their role in the global circulation. From November 2011 through January 2012 DYNAMO/AMIE deployed the NCAR S-PolKa radar on Addu Atoll in the Maldives. This tiny island is only 3 m above sea-level and is completely surrounded by the Indian Ocean. The NCAR S-PolKa radar had both dual-polarimetric and Doppler capabilities. By emitting and receiving horizontally and vertically polarized pulses, this radar calculated additional moments of the drop size distribution that described the physical properties of the particles in each volume of air sampled by the radar. This dissertation interprets these physical properties as an indication of the hydrometeors/microphysical processes in each radar sample volume using a particle identification (PID) algorithm (Vivekanandan et al., 1999) and investigates how they are related to the kinematic structure of observed and simulated MCSs. While the PID algorithm has its limitations, these observational data provide high-resolution microphysical data over large contiguous spatial and temporal scales. The fidelity of the NCAR PID used during DYNAMO/AMIE was deemed accurate by independent assessments conducted in this dissertation and by Martini et al. (2015).

Using an innovative spatial compositing technique to map the location of the hydrometeors identified by the PID with respect to the classic kinematic MCS structure, this dissertation directly shows that hydrometeors were systematically organized with respect to the mesoscale airflow of an MCS during DYNAMO/AMIE. Convective updrafts were characterized by moderate rain within their core since the rapidly rising air was likely preventing the formation of large particles.
Heavy rain and narrow shallow columns of graupel/rimed aggregates were just behind the core, where convective downdrafts were likely present. In general, graupel/rimed aggregates are not expected to reach high altitudes in tropical oceanic environments since their updrafts are relatively weak (e.g. LeMone and Zipser, 1980; Zipser and LeMone, 1980; Mohr and Zipser, 1996; Nesbitt et al., 2000). Wet aggregates occurred behind the convective updraft in a narrow layer near the melting level, where vertical motions were likely less intense. Above the melting level, convective updrafts were dominated by dry aggregates where turbulent motion was expected. The cloud edges were characterized by small ice crystals. Thus, hydrometeors were systematically organized around the mesoscale convective updraft in MCSs during DYNAMO/AMIE in a manner that is entirely consistent with the known dynamical characteristics of MCSs.

Hydrometeors within the stratiform midlevel inflow region were characterized by systematically decreasing rain intensities with distance from the center of the storm and sequential layers of small ice crystals, dry aggregates, and wet aggregates descending from cloud top. The layered frozen hydrometeor pattern has been previously observed (e.g. Houze and Churchill, 1987, Park et al., 2009; Bechini et al., 2013; Jung et al., 2012) and is expected since the stratiform region is characterized by generally weak upward motion aloft, which is not strong enough to suspend hydrometeors and enables particles to gradually fall due to gravitational settling (Houze, 1997). Occasionally, shallow pockets of graupel/rimed aggregates are observed at the top of the wet aggregate layer (as foreseen by Leary and Houze, 1979). While the graupel/rimed aggregates observed in the midlevel inflow region could be left over from previous convection or generated from small-scale isolated embedded convection, it is also possible that strong vertical wind shear along the boundary of the midlevel inflow could create enough turbulence to facilitate the development of these hydrometeors. Graupel/rimed aggregates have been previously observed
within the stratiform precipitation in the equatorial maritime continent (Takahashi and Kuhara, 1992; Takahashi et al., 1995), West Africa (Evaristo et al., 2010; Bouniol et al., 2010), Oklahoma (Zrnić et al., 1993), Taiwan (Jung et al, 2012), and Europe (Hogan et al., 2002). Thus, the midlevel inflow systematically organizes hydrometeors within an MCS in a manner that is consistent with the known dynamical characteristics of the stratiform region.

By framing the locations of the hydrometeors relative to the kinematic structure of MCSs, this dissertation directly expands upon the conceptual model of the structure of MCSs (especially the layered circulation conceptual models of Moncrieff (1992) and Kingsmill and Houze (1999a)). GATE and MONEX, with quantitative three-dimensional reflectivity, provided the research community with details of the basic convective and stratiform sectors of MCSs and of the organization of latent and radiative heating within these storms (Houze, 1982). TOGA COARE, through the deployment of Doppler radar technology, added the kinematic structure to the model. Now, DYNAMO/AMIE, through the dual-polarization radar analysis conducted in this dissertation, adds the hydrometeor structure to the model and continues a 50 year tradition of using evolving radar technology to further our understanding of the fundamental nature of MCSs.

An alternative way of interpreting the PID is to view the categories as an indication of the microphysical process acting in each radar sample volume. This interpretation is very important since it enables the PID data to be directly compared to numerical simulations. When the PID is interpreted in terms of hydrometeors it cannot validate numerical simulations since the hydrometeors definitions used in numerical simulations and the PID are fundamentally different, even though they use similar names. However, this mismatch is resolved by interpreting the PID in terms of microphysical processes. The second part of this dissertation capitalizes on this fact and investigates whether stratiform precipitation simulated by three routinely available
microphysical parameterizations is characterized by a layered ice microphysical pattern that is consistent with DYNAMO/AMIE observations, our theoretical understanding of microphysical processes, and each other. The microphysical parameterizations used in this dissertation include the Milbrandt-Yau Double Moment Scheme (Milbrandt and Yau, 2005a; b), the Morrison 2-Moment Scheme (Morrison et al., 2009), and the WRF Double-Moment 6-Class Scheme (Lim and Hong, 2010).

Given that microphysical processes are inherently linked to the dynamical structure of convection, the observed and simulated ice microphysical patterns can only be equitably compared if the airflow in observations and simulations are the same. In order to satisfy this requirement, while still allowing the circulation to interact and evolve with the microphysics, the simulations in this dissertation assimilated S-PolKa radial velocity data into the Advanced Research version of the Weather Research and Forecasting (WRF-ARW) model. Thus, this methodology constrained simulations to have a kinematic structure consistent with observations but allowed the microphysical processes to evolve freely with minimal manipulation. While this methodology did not force the simulations and observations to have exactly the same convective-scale dynamical structure, differences in these small-scale perturbations did not fundamentally impact the results of this dissertation.

Broadly speaking, the general pattern of ice microphysical processes simulated by each of the three parameterizations was consistent with DYNAMO/AMIE data and previous findings and interpretations put forth by Leary and Houze (1979a), Houze (1981; 1989), and Churchill and Houze (1989). Specifically, deposition occurred anywhere above the 0°C level where upward motion existed, aggregation occurred at and above the 0°C level, riming occurred near the 0°C level, and melting occurred at and below the 0°C level. Additionally, all suggest that riming occurs
within the midlevel inflow region where small-scale vertical velocity perturbations exist. It is unclear whether these velocity perturbations result from small scale turbulence, embedded small-scale convection, or is residual from previous convective cells. Overall, the similarities suggest that these three parameterizations correctly simulated to first-order the mesoscale interactions between the dynamical and ice microphysical processes.

Despite these first-order similarities, substantial differences were found when the details of the simulated ice microphysical pattern were considered. Aggregation and riming in each of these parameterization occurred over too deep of a layer compared to both observations and previous theoretical studies (e.g. Hobbs et al., 1974; Mohr and Zipser, 1996; Nesbitt et al., 2000). Additionally, simulated riming occurred too frequently. These discrepancies may help explain some of the problems that commonly plague numerical simulations. Excessive aggregation may account for the inability for simulations to adequately represent stratiform precipitation by creating ice hydrometeors that are too large and fall out too quickly (e.g. Morrison et al., 2009; Hagos et al., 2014; and others extending back to Fovell and Ogura, 1988). Excessive riming may account for simulated oceanic convection often being too deep, due to too much latent heat release, and containing too much graupel (e.g. Wiedner et al., 2004; McFarquhar et al., 2006; Lang et al., 2007; Li et al., 2008; Lang et al., 2011; Varble et al., 2014).

When the detailed spatial pattern of ice microphysical processes were compared among the parameterizations themselves it was found that no two schemes produced similar patterns for every ice process. These discrepancies in the detailed microphysical pattern among the schemes likely had important implications on the nature of the simulated convection. Despite having the same first-order kinematic and ice microphysical pattern, the horizontal and vertical distributions of reflectivity created by the three parameterizations differed drastically when compared to each other.
and observations. While the spatial pattern of ice microphysical processes was not the only difference among the parameterizations, these spatial ice microphysical patterns must have some impact on the simulated reflectivity pattern since ice processes impact the radiative and latent heating structure of convection. These differences have major implications for forecasting the correct amounts of precipitation and net latent heating by MCSs, and therefore stand as a major area of needed improvement in simulations.

As a whole, the results from this dissertation provide insight into how microphysical processes are related to the mesoscale dynamical structure of MCSs and provides potential reasons why errors within numerical simulations exist. Building upon the classic MCS model developed using newest radar technology available during GATE (quantitative three-dimensional reflectivity) and TOGA COARE (Doppler velocity measurement), the dual-polarimetric radar analysis conducted in this dissertation provides microphysical information that demonstrates that the mesoscale circulation within an MCS is directly related to the organization of hydrometeors and microphysical processes within the storm. The simulations conducted as part of this dissertation using different microphysical parameterizations generally produced ice microphysical patterns within their midlevel inflow region that were similar to each other and to observations. However, discrepancies existed, especially in terms of aggregation and riming.

This understanding of the spatial pattern of microphysical processes within observed and simulated tropical oceanic MCSs begins to provide insight into the latent heating and radiative patterns in MCSs and errors within numerical simulations. For example, the shallow vertical extent of riming witnessed in the dual-polarimetric radar observations in the midlevel inflow region suggests that the latent heat attributable to riming is restricted to a shallow layer just above the 0°C level. This implies that deposition is the primary microphysical process responsible for creating
the upper portion of the top-heavy heating profile in the midlevel inflow region. Each of the three parameterizations in this dissertation produced riming over too deep of a layer. While it is unclear whether this riming discrepancy resulted from errors in the simulated vertical motion or parameterized riming, these modeling results suggest that riming contributed latent heat over too deep of a layer. Thus, the simulated latent heating profiles were likely either the wrong shape or magnitude. If the simulated latent heating profiles were indeed erroneous due to riming, this riming error may help explain why each simulation had a different radar reflectivity pattern (i.e. precipitation pattern) despite having the same first-order kinematic and microphysical patterns. It is anticipated that simulations will not become more accurate and consistent until these discrepancies among the parameterizations, observations, and theory are improved. Thus, future research will need to focus on isolating the specific sources of these microphysical errors within parameterizations and finding their remedy.
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Hannah C. Barnes grew up in suburban Milwaukee, Wisconsin. While she was terrified of thunderstorms as a young child, over time her fear of the weather transformed into fascination. A turning point in this transformation was a sixth grade assignment to plot the track of hurricanes. This project also laid the foundation for Hannah’s passion for tropical meteorology.

Hannah attended the University of Wisconsin – Madison where she majored in atmospheric and oceanic sciences and obtained her Bachelors of Science in 2010. During her undergraduate studies Hannah was given the opportunity to conduct research with Professor Daniel J. Vimont at the University of Wisconsin – Madison and Dr. Jerald Brotzge and Dr. Somer Erickson at the University of Oklahoma as part of the National Weather Center’s Research Experience for Undergraduates (NWC REU) program. The NWC REU provided Hannah with her first exposure to radar data.

Hannah joined Mesoscale Group led by Professor Robert A. Houze Jr. at the University of Washington in 2010. While attending the University of Washington Hannah’s research used spaceborne and ground-based radars to understand the structure and variability of tropical oceanic mesoscale convective systems. In 2011, Hannah traveled to the Maldives to participate in the Dynamics of the Madden-Julian Oscillation / ARM MJO Initiation Experiment (DYNAMO/AMIE). After working beneath NCAR’s S-PolKa radar for six weeks, Hannah’s passion for radar meteorology was undeniable. The experience would be guiding force in her
subsequent research. Hannah received her Masters of Science in 2013 and graduated from the University of Washington with her Doctor of Philosophy in 2016.