Comparison of Observed and Simulated Spatial Patterns of Ice Microphysical Processes in Tropical Oceanic Mesoscale Convective Systems

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Key Points

- Simulated ice processes in the midlevel inflow are consistent with radar observations and theory.
- Of the processes considered, simulated riming and aggregation differ the most from observations.
- Disparate simulated ice microphysical patterns factor into reflectivity differences.

Abstract

To equitably compare the spatial pattern of ice microphysical processes produced by three microphysical parameterizations with each other, observations, and theory, simulations of tropical oceanic mesoscale convective systems (MCSs) in the Weather Research and Forecasting (WRF) model were forced to develop the same mesoscale circulations as observations by assimilating radial velocity data from a Doppler radar. The same general layering of 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, this study is consistent with the layered ice microphysical pattern portrayed in previous conceptual models and indicated by 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 study 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 compared to the observations and theory. 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 model reflectivity patterns.

It is concluded that improved parameterizations of ice-phase microphysics will be essential to obtain reliable, consistent model simulations of tropical oceanic MCSs.

Index terms and Keywords
AGU: 3329 Mesoscale Meteorology, 3371 Tropical Convection, 3374 Tropical Meteorology, 3315 Data Assimilation, 3314 Convective Processes

Author: Mesoscale Convective Systems, Cloud Microphysics, Stratiform Precipitation, Dual-Polarimetric Radar

Text

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 in 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 [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 and Houze* [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. One approach is to use a particle identification algorithm (PID).Traditionally, PIDs have been used to identify the dominant hydrometeor type within convection [e.g. *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 particles, not their exact size or density, they 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/accretion, or riming). This study capitalizes on this capability of the PID methodology.
PID classifications have limitations that must be considered when the data is analyzed. Data provided by PIDs is restricted since the radar statistically samples large volumes that contain many particles and only provides data on the most likely dominant microphysical process. Additionally, the PID data is 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 high resolution 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 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 towards 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 \cite{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 midlevel inflow structure \cite{Kingsmill and Houze, 1999a}. Because of this repeatable kinematic structure, it is possible to composite the pattern of microphysical processes relative to the mesoscale midlevel inflow dominating the stratiform region and the mean steeply rising current in the convective region.

\textit{Barnes and Houze} \cite{2014} took this compositing approach and applied 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 field campaign / ARM MJO Investigation Experiment (DYNAMO/AMIE) \cite{Yoneyama et al., 2013}. Compositing PID data from 36 MCSs with a midlevel inflow, they found that frozen hydrometeors had a systematic layered pattern with small ice crystals near echo top, dry aggregates at midlevels above the $0^\circ$C level, and melting particles near the $0^\circ$C level. Additionally, graupel/rimed aggregates were found occasionally in shallow pockets.
just above the 0°C level. Their results were insensitive to the morphology of the MCS.

Previous studies have found a similar layered frozen hydrometeor structure 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]. However the PID-based composite analysis conducted by Barnes and Houze [2014] was the first study 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 Barnes and Houze [2014] are interpreted in terms of ice microphysical processes, their results indicate that the dominant ice microphysical processes in MCSs over the tropical ocean systematically transition downward from primarily deposition at high levels to a layer of considerable aggregation and some deposition 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], Houze and Churchill [1987], and Braun and Houze [1994]. Given that Barnes and Houze [2014] has demonstrated 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 the performance of 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., 2015].
One aspect of microphysical parameterizations that has been largely neglected in prior intercomparison studies is the spatial distribution of microphysical processes. The spatial distribution of ice microphysical processes is particularly important since Varble et al. [2011] demonstrated that significant differences exist in the mixed and ice phase regions of observations and simulations and Wang et al. [2009] suggested that inconsistencies among observations and simulations result from the treatment of frozen hydrometeors in parameterizations. 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/accretion. 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.
The current study assimilates radial velocity data from 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 full interaction with the microphysical processes. 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 study will explicitly investigate if three different microphysical parameterization schemes produce large-scale spatial patterns of deposition, aggregation/accretion, riming, and melting that are:

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

2. Methodology

2.1 Microphysical Interpretation of Particle Identification (PID) Algorithm

By emitting and receiving horizontally and vertically polarized pulses, dual-polarization 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 Bringi and
Chandrasekar [2001] for a comprehensive description of dual-polarimetric radar variables.) Vivekanandan et al. [1999] developed a technique that classifies regions within convection by combining dual-polarization radar variables and rawinsonde temperature profiles. Included in this frequently used algorithm, which is referred to as a PID, are five frozen hydrometeor categories. Barnes and Houze [2014] referred to these frozen hydrometeor categories as horizontally oriented ice crystals, small ice crystals, dry aggregates, wet aggregates, and graupel/rimed aggregates. (For additional details about the PID used during DYNAMO/AMIE see Rowe and Houze [2014] and Barnes and Houze [2014].) While these categories are named in terms of particle type, the names imply the microphysical processes producing the particles.

Table 1 shows the dual-polarimetric and temperature thresholds used to define each of these categories. One of the defining characteristics of horizontally oriented ice crystals is large ZDR values, which indicate that the major axis of these particles have a strong horizontal component to their orientation. However, both horizontally oriented ice crystals and small ice crystals are associated with the smallest horizontal reflectivity and have large negative values of LDR, which suggests that both particles are the smallest frozen hydrometeors identified by the radar and have an undefinable shape. The most likely way that a frozen particle can grow large enough to be detected by the radar and not have the characteristics of an aggregate of particles is through deposition. Barnes and Houze [2014] found that horizontally oriented ice crystals during DYNAMO/AMIE were often observed at temperatures between -10°C and -20°C and in regions of vertical wind shear. Mason [1971] and Hobbs [1974] suggest that vertical wind shear at these temperatures promotes enhanced depositional growth. Both horizontally oriented ice
crystals and small ice crystals could have been advected from another part of the storm, but even in that case any subsequent growth of these advected particles would most likely result from deposition. Compared to the horizontally oriented ice crystals and small ice crystals, dry aggregates have higher reflectivity and their LDR values are less negative. Thus, dry aggregates are larger than horizontally oriented ice crystals and small ice crystals and have a more uniform shape [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. 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. This interpretation of the graupel/rimed aggregates assumes that these particles are formed locally, not advected from other regions of the storm. This is a valid assumption since the authors have analyzed the dual-polarimetric variables associated with dozens of MCSs with a midlevel inflow and graupel/rimed aggregates were always restricted to a narrow layer just above the 0°C level. If these rimed particles had been advected from a more
convectively active portion of the storm, we would expect to see graupel/rimed aggregates at higher altitudes. Using these interpretations the PID categories can be used to map the spatial pattern of ice microphysical processes within a region of radar echo in an MCS. Specifically, horizontally oriented ice crystals and 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 occur 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 towards 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 (Table 1) and because few validation studies exist. Vivekanadan et al. [1999] pointed out that the complex relationships used in the PID are unavoidable and the soft boundaries used in the fuzzy-logic algorithm provides one of the best ways to handle these relationships.

We do not attempt here to address all the uncertainties in PID algorithms. We are 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, *Barnes and Houze* [2014] and *Martini et al.* [2015] have analyzed the performance of the PID during DYNAMO/AMIE and concluded that it was accurate. We are therefore confident that the DYNAMO/AMIE PID dataset describes the dominant ice microphysical process with sufficient accuracy for the purposes of this study.

Our aim 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 fundamentally different. In numerical simulations, hydrometeors are defined based on the particle’s size and density. The PID defines hydrometeors based on their physical characteristics including 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, PID data can be interpreted in terms of the microphysical processes producing the particles, and numerical simulations directly calculate the same processes. Our comparisons therefore are based on processes, not mixing ratios.

### 2.2 Classification of Microphysical Processes in WRF

Microphysical parameterizations within WRF only routinely output the hydrometeor mixing ratios and, if the scheme is double-moment, number concentrations. However, dozens of variables describe the interactions among the mixing ratios of individual hydrometeor types. By grouping these hydrometeor mass interaction variables in terms of the microphysical process they describe, additional three-dimensional fields depicting the mass production rate of microphysical processes can be output. This study 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 mass interaction variable that described how a frozen hydrometeor collected water vapor.
- Aggregation included any hydrometeor mass 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. Thus our definition of aggregation includes all forms of ice accretion. Variables describing how aggregation impacts the number concentration of these particles were not included. For purposes of this study, the impact of this exclusion is relatively minor and will be discussed below.
- Riming included any hydrometeor mass 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 mass interaction variable that described how a frozen hydrometeor became a liquid hydrometeor.

The current study tests three routinely available microphysical parameterization schemes 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 2 shows how the hydrometeor mass interaction variables in each parameterization were grouped into microphysical processes.

2.3 Data Assimilation Simulations

One of the most challenging aspects of comparing observed and simulated 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 methodology provides a straight-forward method to equitably compare parameterizations, it does not address the fact that the microphysics evolve in concert with the dynamics of the convection.

Our study uses data assimilation to require simulations to develop a mesoscale circulation similar to observations. By so doing, we enable the microphysical processes to evolve freely but only in concert with a dynamically accurate simulation. The microphysics can then be compared to dual polarization radar observations with minimal contamination from erroneous 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 investigating 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 influence 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 is the most appropriate radar variable to assimilate in this study since it forces simulations to have similar mesoscale dynamical evolution while impacting the microphysical processes as little as possible.

Radial velocity data used in this study 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 minutes. (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 study 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 study 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 study is provided in Table 3 and the nested domains with 3 and 1 km resolution, respectively, are shown in Figure 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 completed in the inner domain with 1 km resolution. The inner domain is located mostly east of the radar because the scans 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 seconds and data was output every 15 minutes. This model output frequency is appropriate given that the current study is only interested in the spatial pattern of the microphysical processes, not the temporal evolution of the microphysical pattern.

The WRF-based ensemble Kalman filter (EnKF) system used in this study is same as Meng and Zhang [2008a; b] and Zhang et al. [2009]. Details about the EnKF architecture are provided in Table 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. 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.

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, we 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 vertical cross sections 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). This analysis methodology has been applied to different time steps and has been used to composite the same cross section at multiple times. In all cases the composites are fundamentally the same (not shown). 1930 UTC 23 December and 1445 UTC 16 October were specifically selected since they provided the largest sample size.

The midlevel inflow observed by S-PolKa on 23 December descended predominantly from west to east. On 16 October the midlevel inflow observed by S-PolKa 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, 6 cross sections were taken from each ensemble member during the 16 October simulations and 5 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 study.

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 uniform height coordinates.

2. Isolate 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 was 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 was ignored. The threshold changed due to convective intensity differences between 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 between 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 is 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 we previously converted the vertical coordinate of the cross section 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. It is appropriate to use a best fit linear line since we previously converted the vertical coordinate of the cross section from unevenly spaced eta values to uniform height intervals.

8. *Find the base of the core of the midlevel inflow.* Find the maximum point with 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 was arbitrary defined. Results were insensitive to the length scale. Changing the length scale only changed 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/or 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 was arbitrary defined. Results were insensitive to this length scale. Changing this length scale only changed 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 horizontally 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 as 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 were 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 study 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 mass 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 in this study was represented in terms of their composite frequency. At each grid point the number of cores that had a nonzero microphysical mass production rate was normalized by the total number of cores. 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 particles account for the largest radar return power. The
dominant microphysical process in WRF is defined based on the mass production rate of
the process, which cannot be observed using the PID. 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 study indicate where processes were statistically more and less likely to occur
relative to the midlevel inflow, which is directly comparable to the data provided by the
PID.

This study also generated 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 study 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. Dual-Polarization Radar Observations of Mesoscale Convective Systems

*Barnes and Houze* [2014] identified 36 MCSs during DYNAMO/AMIE that
contained a midlevel inflow and created spatial composites that showed that the
DYNAMO/AMIE PID data had a systematic spatial pattern relative to the midlevel
inflow. Squall line and non-squall MCSs were composited separately in *Barnes and
Houze* [2014] 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. The current study simulates a squall line and non-squall MCS that were included in the composites in *Barnes and Houze* [2014]. It is important that this study considers two different types of MCSs in order to investigate if the morphological structure of convection impacts the simulated spatial pattern of ice processes.

The squall line simulated in this study occurred on 23 December 2011 (Figure 2a-2c). This case was selected since it was one 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 2a). Stratiform precipitation extended from the convective line westward for 100-200 km. Figure 2b and 2c shows a vertical cross section of reflectivity and radial velocity data along the red line in Figure 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.

Five vertical cross sections were taken through the 23 December squall line and the PID data from these cross sections were spatially composited around the observed midlevel inflow using a methodology directly analogous to the method described in Section 2.4 (For a complete description of the spatial compositing technique applied to
the PID see Barnes and Houze [2014]). The shading in Figures 3a-3e show the frequency in which horizontally oriented ice crystals, small ice crystals, dry aggregates, graupel/rimed aggregates, and wet aggregates occurred at a given location relative to the midlevel inflow. The dotted magenta line represents the approximate location of the midlevel inflow. These PID composites 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. The isolated region of horizontally oriented ice crystals suggests that deposition was enhanced just above the midlevel inflow near the anvil end of the midlevel inflow, where the shear and temperature conditions correspond to those identified by Mason [1971] and Hobbs [1974] as favoring deposition. Graupel/rimed aggregates were located in a 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 were observed within one of the 36 midlevel inflow regions investigated in Barnes and Houze [2014] these hydrometeors always occurred in shallow, isolated pockets. Additionally, Rowe and Houze [2014] analyzed all MCSs observed during DYNAMO/AMIE and also found that graupel/rimed aggregates in stratiform regions were confined to a narrow region above the 0°C level. Thus, the hydrometeor pattern observed during the 23 December squall line is entirely consistent with the systematic pattern identified in Barnes and Houze [2014] and is long thought to exist in oceanic tropical convective systems [Leary and Houze, 1979b; Houze, 1989].

Interpreting the PID in terms of processes suggests that ice processes on 23 December were layered with only depositional growth near echo top, aggregation at
midlevels above the 0°C level, and melting and shallow pockets of riming 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. We therefore interpret the layer where aggregation is occurring as a zone where both aggregation and vapor deposition were active. This interpretation is important when comparing the PID with model output since the parameterizations 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 non-squall MCS simulated in this study persisted over the S-PolKa radar domain for nearly 18 h on 16 October 2011. This MCS was selected since it was one the largest MCSs observed by S-PolKa during DYNAMO/AMIE. Figures 2d-2f 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 have been conducted when the 16 October MCS was in the earlier stages of its lifecycle. The convective and stratiform regions were organized less systematically with Figure 2d showing pockets of high reflectivity convective cores scattered within an expansive region of weaker reflectivity stratiform precipitation. Figures 2e and 2f show a cross section of reflectivity and radial velocity data through a portion of the stratiform precipitation along the red line in Figure 2d. Five cross sections were taken within southeast portion of this storm and PID data was spatially composited around the
midlevel inflow (Figure 3f-3j) in a manner analogous to the methodology described in Section 2.4. (For a complete description of the spatial compositing technique applied to the PID see Barnes and Houze [2014]).

While some hydrometeor/microphysical differences existed between the 23 December squall line and 16 October non-squall MCS, the composited DYNAMO/AMIE PID data had the same fundamental spatial pattern relative to the midlevel inflow in both cases. Similar to 23 December, the 16 October non-squall MCS had a well-defined midlevel inflow (Figure 2c, 2f). 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. Additionally, the 16 October non-squall MCS was shallower than the 23 December squall line (Figures 2b, 2e) and its precipitation was weaker. A weak brightband was apparent on 16 October, but a portion of the precipitation did not reach the surface. Loops of reflectivity and radial velocity (not shown) suggest that the midlevel inflow on 16 October weakened the MCS, instead of forcing the MCS as observed on 23 December. A few differences in the detailed spatial patterns of the ice processes also exist 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, had a shallower dry aggregate layer, and more prevalent horizontally oriented ice crystals. This comparison suggests that the stratiform region associated with the 16 October non-squall MCS lacked riming and had less aggregation but had stronger deposition than the 23 December squall line. While these differences may be attributable to weaker vertical motions on 16 October than on 23 December, observational data to verify this theory is unavailable. Differences in the prevalence of horizontally oriented ice crystals between these MCSs have little impact on
the conclusions of the current study since both horizontally oriented ice crystals and
small ice crystals are predominantly associated with deposition. The combination of these
two hydrometeors indicates that deposition is the predominant microphysical process
near echo top in each MCS. Thus, while differences in the spatial microphysical pattern
existed between these MCSs, these differences were relatively minor. Overall, 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 in a manner that was consistent with *Barnes and Houze* [2014].

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 4 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, the vertical velocity component is
often an order of magnitude less than the horizontal velocity component within the
midlevel inflow region. Streamlines calculated within the plane of the WRF cross
sections are oriented nearly horizontal (not shown). Additionally, 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). This underestimation by S-PolKa does not impede this study
since validating the absolute magnitude of the simulated midlevel inflow is not our
objective. We only require 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 4 shows that each ensemble had a midlevel inflow whose slope and magnitude was similar to the S-PolKa observations.

The shading in Figure 5 shows the composite 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 studies [Houze 2004; 2014]. Note that the color bars in Figure 5 have different scales. The scale used for the 23 December squall line is twice as large as the scale used for the 16 October non-squall MCS. 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. The magnitude of these vertical velocities cannot be validated since in situ vertical velocity data was unavailable during these storms.

The small-scale vertical velocity differences between the parameterizations in Figure 5 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, this study proceeds and compares the observed and simulated spatial pattern of ice processes within the midlevel inflow region 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.
5. Microphysical Comparison

5.1 Deposition

Figure 6 shows the frequency of simulated deposition within the midlevel inflow region 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 and composites presented in Barnes and Houze [2014], which show a layer of dry aggregates and graupel/rimed aggregates topped by a layer of horizontally oriented ice crystals and 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 indicate 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 always everywhere above an altitude of 5 km, 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 6). Because upward motion was statistically more frequent above the midlevel inflow and near echo top, deposition was more common near 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 5). 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. Houze, 1981; Gamache and Houze, 1983; Houze and Churchill, 1987; Houze, 1989; Braun and Houze, 1994].

Overall, deposition occurred more frequently in each of the three parameterizations on 16 October than 23 December. This difference is consistent with the PID composites. Recall that horizontally oriented ice crystals are indicative of regions of enhanced deposition and these particles were more prevalent on 16 October than 23 December (Figures 3a and 3f). These simulations suggest that the differences in the prevalence of horizontally oriented ice crystals between these MCSs may be related to vertical velocity differences. Specifically, deposition may have been more enhanced on 16 October than 23 December since vertical motions on 16 October were more horizontally uniform and had less small-scale variability within the -10°C to -20°C temperature range (Figure 5) favorable for enhanced deposition [e.g. Mason, 1971, Hobbs, 1974]. Vertical velocity is expected to differ in these cases. While the MCS on 16 October was in a later stages of its lifecycle and had midlevel inflow that remained nearly parallel with the surface, the MCS on 23 December was an active squall line with a steeply sloping midlevel inflow (Figure 2c and 2f)

5.2 Aggregation

The composite frequency of simulated aggregation on 23 December and 16 October is shown in shading in Figure 7. Aggregation had the most consistent spatial pattern considered in this study with all parameterizations producing aggregation
everywhere above the 0°C level. While MOR had a slightly reduced aggregation frequency 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. Note that this study does not include the variables within MY and MOR that change the number concentration of snow due to aggregation. While these number concentration variables are important terms in the parameterizations, including them would not have changed our results or conclusions since our current definition already shows that aggregation always occurs wherever temperatures are below 0°C.

While the microphysical parameterizations were consistent with the PID composites 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 reaching echo top and occurring at temperatures below -20°C. Figure 3c and 3h indicates that aggregation never extended to echo top on 23 December or 16 October. The relatively shallow extent of aggregation is a robust pattern in the PID data that was observed in each of the 36 midlevel inflows analyzed in Barnes and Houze [2014]. Studies such as Houze and Churchill [1987] and Braun and Houze [1994] indicate that aggregation is most common within 1-2 km of the melting level in the stratiform portion of an MCS. Additionally, this simulated pattern disagrees with laboratory and observational studies that have demonstrated that aggregation rarely occurs below -20°C [e.g. Hobbs et al., 1974].

One factor in this altitude discrepancy between the simulations and PID may be related to the fact that simulated aggregation is defined in this study 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 was reduced and deep MCSs with stratiform precipitation was increased when snow and ice were prevented from accreting onto graupel in simulations that used a single-moment microphysical parameterization. Recall that these accretion variables were included within the aggregation term used in this study. Thus, the parameterization of aggregation in these schemes may be a factor limiting the ability to accurately simulate stratiform precipitation. This deficiency in stratiform precipitation is a problem that has been discussed in several modeling studies including Morrison et al. [2009], Hagos et al. [2014], and others extending back to Fovell and Ogura [1988]. Evidence of insufficient stratiform precipitation in our simulations will be presented in section 6.

5.3 Rimming

Figure 8 shows the frequency of simulated 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 that occasionally occurred at altitudes as great as 10
km. It is important to keep in mind that simulated riming in this study 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 (i.e. the approximate location of the
0°C level).

Despite these broad similarities between these two MCSs, riming was the process
analyzed in this study that displayed some of the largest differences among individual
parameterizations and between the MCSs. A striking difference between the schemes in
both cases was that MY sometimes produced riming well below an altitude of 5 km.
Figure 8a indicates that the 23 December MY simulations had near surface riming in
nearly 50% of the cross sections. This near surface riming 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
along the equator. While our definition of riming includes processes that could result in
water-coated frozen hydrometeors, the frozen core of these hydrometeors are still
expected to melt rapidly and should not be present well below the 0°C level. Thus, the
near surface riming in MY is erroneous. While Figures 8a and 8d suggest riming
simulated by MY occurred below the 0°C level, the rate of this riming was very low (not
shown). Thus, this behavior is a model artifact that does not substantially affect the
results of this study and likely does not significantly impact the structure of the simulated
MCS.

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 8). 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 at the anvil end of the midlevel inflow. Rimming in MY and WD was frequently deep near the convective end and was concentrated into a narrow layer near the anvil end. Peak riming in MOR was concentrated in the convective half of the midlevel inflow and was relatively rare in the anvil half.

Rimming occurred within a shallower layer and had a more consistent spatial pattern on 16 October (bottom row of Figure 8). Each parameterization on 16 October had riming concentrated to locations within a layer near the 0°C level. Unlike the 23 December squall line, simulations of the 16 October MCS indicated that the frequency of riming was similar near the convective and anvil end of the midlevel inflow. These differences between the cases are expected since the 16 October non-squall MCS was in a later stage of its lifecycle and had less intense, more horizontally uniform vertical motions than the 23 December active squall line (Figure 5).

Similar to the PID composites, each parameterization indicated that riming sometimes occurred near the 0°C level within the midlevel inflow. Leary and Houze [1979b] were the first to discuss the possible presence of rimed particles in the stratiform region of MCSs. They inferred the presence of riming in this layer indirectly from drop-size distributions measured below the brightband. Since then observational and modeling studies have identified rimed particles within the stratiform precipitation over the Indian Ocean [Suzuki et al., 2006; Rowe and Houze, 2013], equatorial maritime continent [Takahashi and Kuhara, 1992; Takahashi et al., 1995], West Africa [Evaristo et al.,
It is a positive result that all three microphysical schemes identify this riming layer.

Riming requires supercooled water, which is produced through upward motion or turbulence. Barnes and Houze [2014] suggested that the riming within the midlevel inflow region during DYNAMO/AMIE could have resulted from shear-induced turbulence, small embedded convective cells, or residual upward motion left from a more convective phase of the precipitation. Each of these potential sources of riming would result in pockets of supercooled water, and associated riming, intermittent on space and time scales smaller than the mesoscale midlevel inflow current. However, Barnes and Houze [2014] could not confirm the association between convective-scale vertical velocity perturbations and riming since they lacked vertical velocity observations. Additionally, it remains possible, though unlikely, that the rimed particles observed in the PID data could have been advected into the midlevel inflow region from a more convectively active portion of the storm. The exact sources of the rimed particles is a topic beyond the scope of the present study.

Even though assimilation has forced our simulations to have a similar mesoscale vertical velocity pattern, convective-scale vertical velocity differences exist. In each simulation parameterized riming often was not strongly correlated with the mesoscale vertical velocity (orange contours in Figure 8). The fact that this behavior occurred in every simulation analyzed in this study provides strong evidence that the riming is associated with convective-scale vertical velocity perturbations superimposed on the midlevel inflow. The current study cannot determine whether the convective-scale
velocity perturbations necessary for riming are shear or buoyancy induced. However, our results confirm the inference of Barnes and Houze [2014] that riming within the midlevel inflow region during DYNAMO/AMIE was most likely the result of convective-scale vertical velocity perturbations.

While riming is present in both in observations and simulations, there are discrepancies between observed and simulated composites in terms of the detailed spatial pattern of riming. The most systematic and significant of these discrepancies is the frequency of riming and the depth of the riming layer. The bottom row of Figure 8 indicates that riming could occur at any given location along the midlevel inflow in the 16 October case with a frequency of at least 0.7 in each parameterization. This high frequency strongly contrasts the observational PID. During the later stages of the 16 October MCS, PID observations indicated that riming was absent from the stratiform portion of the MCS (Figure 3i). Riming was observed during the 23 December squall line, but the riming frequency peaked at 0.6 (Figure 3d). PID composites presented in Barnes and Houze [2014] (their Figure 10), which were generated from cross sections using PID data from 36 individual MCSs, indicate that the frequency of riming at a given location within the midlevel inflow region during DYNAMO/AMIE was only 0.2. 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.

Nearly 70% of the cores in each scheme on 23 December had riming at an altitude of 8 km (top row Figure 8). Figure 3d indicates that riming was detected only in very
shallow pockets within the midlevel inflow during the 23 December squall line. These shallow pockets of rime are not a feature unique to the 23 December squall line. Rowe and Houze [2014] and Barnes and Houze [2014] found that all instances of graupel/rimed aggregates (i.e. rime) within the stratiform portion of MCSs were concentrated into a shallow region near the 0°C level. While the PID may be unable identify instances of very weak rime, previous studies suggest that rime should not extend over deep layers within the midlevel inflow region of an oceanic MCS (e.g. Leary and Houze, 1979b). Additionally, rime is expected to be shallow in these storms since these MCSs 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]. Even in convective cores, DYNAMO/AMIE PID data indicated that rime particles only reached a few kilometers above the 0°C level [Rowe and Houze, 2014; Barnes and Houze, 2014]. Thus, we infer that simulated rime in these three parameterizations extends over too deep of a layer. This excessive rime may create too much graupel and cause too much latent heat release, which may contribute to the tendency for simulated oceanic convection in previous studies to be too strong and contain 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].

5.4 Melting

The frequency of simulated melting on 23 December and 16 October is shown in Figure 9. In each set of simulations, these parameterizations indicated that melting occurred near the 0°C (bottom orange line in Figure 9). 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, we see that 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 paper and likely does not significantly impact the structure of the
simulated MCS.

Excluding MY, the simulated spatial pattern of melting was the most consistent
with the PID composites and previous studies of any process analyzed in this study. 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 melting in MOR was slightly more accurate than the
somewhat deeper layer in the WD simulation.

6. Impact of Microphysical Differences

From the foregoing discussions, it is evident that no single scheme accurately
portrays the precise spatial distribution of microphysical processes in either of the two
MCSs considered in this study. 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 10 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 11 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 an anvil in Figures 10b and 10d. However, S-PolKa data does not indicate an anvil (Figure 10a). The reflectivity near the convective end of the midlevel inflow was stronger in MY (Figure 10b) and the brightband was more distinct in WD (Figure 10d). Figures 11b and 11d shows that WD produced a better-defined leading convective line 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 lacked an anvil and had the most distinct convective, transition, and stratiform regions, making it the most consistent with S-PolKa observations (Figure 10c and 10a). Additionally, the horizontal distribution of reflectivity in MOR was most similar to S-PolKa on 23 December (Figure 11c and 11a).

The parameterizations had even more difficulty representing the reflectivity structure on 16 October. As in the 23 December simulations, the MOR simulation on 16 October was most similar to S-PolKa observations, with a well-defined bright band
(Figure 10g and 10e) and isolated convective pockets whose locations were similar to observations (Figure 11g and 11e). However, reflectivity in MOR on 16 October was too high, especially in the anvil end of the composites (right side of Figure 10g) and near the surface in the convective end (left side of Figure 10g). Additionally, the horizontal map of reflectivity in Figure 11g indicates reflectivity in the convective pockets on 16 October tended to be too high in MOR. MY lacked a brightband on 16 October and reflectivity in the anvil end of the composites was often too high (Figure 10f). The horizontal map of reflectivity at the 5 km level in Figure 11f indicates that reflectivity simulated by MY was often too low, reflecting the fact that the brightband in MY was too weak. While the horizontal distribution of reflectivity from WD on 16 October was similar to S-PolKa (Figure 11h and 11e), its reflectivity cross section was drastically different than S-PolKa and the other schemes. Figure 10h shows that WD had a strong brightband across the entire midlevel inflow region and reflectivity was not observed below 3 km. Thus, WD was extremely deficient in stratiform precipitation on 16 October. 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 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 10 and 11 indicate that large reflectivity discrepancies were witnessed among the parameterizations even though data assimilation forced each ensemble to have similar mesoscale circulations and their 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. While these assumed parameter differences contribute to variations in the simulated spatial distribution of precipitation in both the horizontal and vertical dimensions, the details of the spatial distribution of the ice-phase microphysical process must also factor into obtaining the correct precipitation distribution. 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. 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].

7. Conclusions

Barnes and Houze [2014] 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 horizontally oriented ice crystals/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 categories identified as horizontally oriented ice crystals and 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 of Barnes and Houze [2014] therefore indicate that these basic ice-phase microphysical process are distributed in the midlevel inflow region of tropical, oceanic MCSs in a systematic layered pattern.
The current study investigates whether the spatial pattern of ice processes 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 Barnes and Houze [2014], our theoretical understanding of microphysical processes, and each other. The microphysical parameterizations included in the current study 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 fundamentally 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. Two sets of simulations were done to test if the simulated spatial pattern was 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 ice 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 rime near the 0°C level.

- The spatial pattern of the ice microphysical processes was not fundamentally impacted by whether or not the MCS was organized into a squall line.
- Each parameterization had deposition located above the 0°C level wherever upward motion existed. The PID indicated ice particles (horizontally oriented ice crystals, small ice crystals, dry aggregates, or graupel/rimed aggregates) existed everywhere above the 0°C level.
- The small-scale spatial variability of simulated riming is likely caused by differences in the convective-scale vertical velocity motions in the simulations, which supports the theory put forth by Barnes and Houze [2014] that the riming during DYNAMO/AMIE occurred in pockets within the midlevel inflow region due to vertical motion from small-scale shear induced turbulence or embedded convective cells or left as residual from a more convective phase of the storm.

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

- None of the schemes produced similar details in every microphysical pattern considered in this study. 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 existed everywhere above the 0°C
level, regardless of the life stage of the MCS. PID data indicated that aggregation
never occurred near echo top and extended only a short distance above the 0°C level,
especially in MCSs at a later stage of development. Simulated aggregation was
notably inconsistent with previous observational and laboratory studies, which show
that aggregation is nearly impossible below -20°C [e.g. Hobbs et al., 1974].

Aggregation simulated at these very low temperatures could create biases in the
radiative flux and be inimical to the development of stratiform precipitation [e.g.
Varble et al., 2011; Van Weverberg 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 later developmental stages of MCSs, 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 study to include processes that create water-coated frozen
hydrometeors, which may partially account for the high riming frequency near the
0°C level in these simulations.

- Rimming and melting in MY were sometimes located well below the 0°C level, which
is incorrect; frozen hydrometeors never occur near the surface over equatorial oceans.
However, the rate at which these near surface processes occurred in the model was relatively slight and likely do not significantly impact the structure of the MCS. These erroneous near surface rates are 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 composites in Figure 3 and Barnes and Houze [2014]. 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 study generally support the layered pattern presented in the PID and these previous 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 [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; Barnes and Houze, 2014]. While the detailed riming pattern varied among each parameterization used in this study, there was a key similarity among the parameterizations. Riming often occurred outside of the region of large-scale upward motion associated with the midlevel inflow. While the assimilation of radial velocity had produced a similar midlevel inflow in each simulation, convective-scale vertical velocity differences existed among the simulations. Thus, the spatial riming patterns presented in this study provides strong evidence that riming within the midlevel inflow region of these tropical oceanic MCSs was associated with
convective-scale velocity perturbations. Whether the vertical velocity perturbations are shear or buoyancy induced remains to be determined.

This study has shown that even while data assimilation produces simulations that evolve 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 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.

Acknowledgements

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graphics, and two anonymous reviewers who provided helpful comments that improved this study.

References


Zipser, E. J (2003), Some views on “Hot Towers” after 50 years of tropical field programs and two years of TRMM data, *Meteor. Monogr.*, 29(51), 49-58.


Table 1: Approximate range of values used in the NCAR S-PolKa PID during DYNAMO/AMIE.

<table>
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<tr>
<th>Hydrometeor Name</th>
<th>Microphysical Process</th>
<th>$Z_H$ (dBZ)</th>
<th>ZDR (dB)</th>
<th>$L_{DR}$ (dB)</th>
<th>$K_{DP}$ (° km$^{-1}$)</th>
<th>$\rho_{HV}$</th>
<th>$T$ (°C)</th>
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<td>Horizontally Oriented Ice Crystals</td>
<td>Deposition</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>
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<tr>
<td>Small Ice Crystals</td>
<td>Deposition</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>Dry Aggregates</td>
<td>Aggregation and Deposition</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>
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<tr>
<td>Graupel/Rimed Aggregates</td>
<td>Riming and Deposition</td>
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<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>
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<tr>
<td>Wet Aggregates</td>
<td>Melting</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>
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Table 2: The definition of the ice microphysical processes adopted in this study and the variables from each parameterization used to define each process.

<table>
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<th>Ice Microphysical Processes</th>
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<td>Aggregation</td>
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<td>Frozen hydrometeors</td>
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<td>collecting other</td>
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<td>(MY)</td>
<td>QCLilh</td>
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<td>Morrison (MOR)</td>
<td>prai, prci</td>
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Table 3: WRF and EnKF setup for simulations.

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Figure Captions

Figure 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.

Figure 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 - f) Same as (a – c) except at 1450 UTC on 16 October 2011.

Figure 3: (a) Composite showing the location of horizontally oriented ice crystals in the midlevel inflow region of the 23 December squall line. Shading represents the frequency of horizontally oriented ice crystals at that location relative to the midlevel inflow. The position of the midlevel inflow is approximated by the dashed diagonal magenta line. The boundaries of the midlevel inflow are marked by black vertical lines. The height axis is normalized so that the melting level is located at 0. (b) Same as (a) except for small ice crystals. (c) Same as (a) except for dry aggregates. (d) Same as (a) except for graupel/rimed aggregates. (e) Same as (a) except for wet aggregates. (f-j) Same as (a-e) except for the 16 October non-squall MCS.

Figure 4: (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.

Figure 5: (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 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top green lines, respectively. 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. (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.

Figure 6: (a) Cross section of the composite frequency of deposition simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average upward motion in m s$^{-1}$ is shown in orange contours with upward motion contoured is every 0.1 m s$^{-1}$ from 0.1 m s$^{-1}$ to 0.5 m s$^{-1}$. 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. (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.

Figure 7: (a) Cross section of the composite frequency of aggregation simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top orange lines, respectively. The composite average reflectivity is shown in gold contours with reflectivity values contoured every 5 dBZ from 5 to 45 dBZ. The edges of the core of the midlevel inflow 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. (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.

Figure 8: (a) Cross section of the composite frequency of riming simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average upward motion in m s\(^{-1}\) is shown in orange contours with upward motion contoured is every 0.1 m s\(^{-1}\) from 0.1 m s\(^{-1}\) to 0.5 m s\(^{-1}\). 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. (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.

Figure 9: (a) Cross section of the composite frequency of melting simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top orange lines, respectively. The composite average reflectivity is shown in gold contours with reflectivity values contoured every 5 dBZ from 5 to 45 dBZ. The edges of the core of the midlevel inflow 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. (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.

Figure 10: (a) Cross section of the average S-PolKa radar reflectivity at 1930 UTC on 23 December 2011. (b) Cross section of the composite average horizontal radar reflectivity simulated by the Milbrandt-Yau (MY) scheme at 1930 UTC on 23 December 2011 in shading. 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.

Figure 11: (a) Horizontal map of S-PolKa reflectivity at an altitude of 5 km at 1930 UTC on 23 December 2011. The red dashed lines indicate where the cross sections shown in Figure 3a and 9a are taken. (b) Horizontal map of reflectivity simulated by the Milbrandt-Yau (MY) scheme at an altitude of 5 km at 1930 UTC on 23 December 2011. The dashed red lines show where the cross sections analyzed in this study were taken. (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 at 1445 UTC at 1445 UTC on 16 October 2011.
**Figure 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.
Figure 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.
Figure 3: (a) Composite showing the location of horizontally oriented ice crystals in the midlevel inflow region of the 23 December squall line. Shading represents the frequency of horizontally oriented ice crystals at that location relative to the midlevel inflow. The position of the midlevel
inflow is approximated by the dashed diagonal magenta line. The boundaries of the midlevel inflow are marked by black vertical lines. The height axis is normalized so that the melting level is located at 0. (b) Same as (a) except for small ice crystals. (c) Same as (a) except for dry aggregates. (d) Same as (a) except for graupel/rimed aggregates. (e) Same as (a) except for wet aggregates. (f-j) Same as (a-e) except for the 16 October non-squall MCS.
Figure 4: (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.
Figure 5: (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 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top green lines, respectively. 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. (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.
Figure 6: (a) Cross section of the composite frequency of deposition simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average upward motion in m s$^{-1}$ is shown in orange contours with upward motion contoured is every 0.1 m s$^{-1}$ from 0.1 m s$^{-1}$ to 0.5 m s$^{-1}$. 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. (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.
Figure 7: (a) Cross section of the composite frequency of aggregation simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top orange lines, respectively. The composite average reflectivity is shown in gold contours with reflectivity values contoured every 5 dBZ from 5 to 45 dBZ. The edges of the core of the midlevel inflow 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. (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.
Figure 8: (a) Cross section of the composite frequency of rime simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average upward motion in m s\(^{-1}\) is shown in orange contours with upward motion contoured is every 0.1 m s\(^{-1}\) from 0.1 m s\(^{-1}\) to 0.5 m s\(^{-1}\). 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. (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.
Figure 9: (a) Cross section of the composite frequency of melting simulated by the Milbrandt-Yau (MY) scheme within the midlevel inflow core at 1930 UTC on 23 December 2011 in shading. The composite average 0°C, -20°C, and -40°C contours are shown in bottom, middle, and top orange lines, respectively. The composite average reflectivity is shown in gold contours with reflectivity values contoured every 5 dBZ from 5 to 45 dBZ. The edges of the core of the midlevel inflow 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. (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.
Figure 10: (a) Cross section of the average S-PolKa radar reflectivity at 1930 UTC on 23 December 2011. (b) Cross section of the composite average horizontal radar reflectivity simulated by the Milbrandt-Yau (MY) scheme at 1930 UTC on 23 December 2011 in shading. 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.
Figure 11: (a) Horizontal map of S-PolKa reflectivity at an altitude of 5 km at 1930 UTC on 23 December 2011. The red dashed lines indicate where the cross sections shown in Figure 3a and 9a are taken. (b) Horizontal map of reflectivity simulated by the Milbrandt-Yau (MY) scheme at an altitude of 5 km at 1930 UTC on 23 December 2011. The dashed red lines show where the cross sections analyzed in this study were taken. (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 at 1445 UTC at 1445 UTC on 16 October 2011.