Microphysical and Near-Storm Environmental control on the Maintenance of Nocturnal Mesoscale Convective Systems: A Case Study

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ABSTRACT

In this study, the sustenance of the nocturnal CIs within mesoscale convective systems (MCSs) developed on 15 July 2015 during the Plains Elevated Convection at Night (PECAN) field campaign were investigated with a combination of in-situ observations and a set of Weather Research and Forecasting (WRF) experiments.

Observational analyses revealed that systems with a greater percentage of CIs near the system edge had greater maintainability than system where CIs tended to cluster in system rear. Two hypotheses were proposed to explain this phenomenon: (a) environmental instability near the system edge CIs were greater due to enhanced moisture above the boundary layer and (b) the kinematic-microphysical structures of systems with system edge CIs evolved in a manner that was favorable for system maintenance. Specifically, dual-polarimetric observations indicate stronger, more extended rear-inflow jet (RIJ) and increased riming growth within the convective updrafts for these systems.

A set of microphysical sensitivity experiments were performed to evaluate the two hypotheses. Since the ambient environmental instabilities were similar between the experiments, internal processes would play a dominate role if significant inter-model differences in updraft strength were found.

Statistical analyses suggest that simulated systems were stronger when rimed particles can sediment at different terminal velocities with regard to their sizes. RIJs in these systems tended to the stronger and more horizontally expanded, allowing more system edge CIs. In these experiments, preferential sedimentation of melting graupel increased the buoyancy gradient near system edge and created stronger negative buoyancy pressure perturbation, which enhanced the system RIJs. Stronger and more horizontally extended RIJs could subsequently strengthen the system by extra riming and deposition when the RIJs transported the graupel back to the updrafts.

Keywords: Precipitation Process; Cloud Microphysics; Mesoscale Dynamics

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Chinese Abstract

本研究利用於 2015 年 7 月 15 日於美國堪薩斯州進行之 PECAN（Plains Elevated Convection at Night）觀測實驗相關資料及 WRF 敏感度實驗研究美國中西部中尺度對流系統在晚間強度維持的原因。觀測資料顯示於晚上減弱的系統有較多對流於系統後方形成（System rear Cls），而於晚上增強的系統則有更多對流於系統邊緣形成（System edge Cls）。利用觀測資料，我們提出兩個不同的假設來解釋這種差異：（一）System edge Cls 附近大氣邊界層上方水氣含量較多和（二）兩種對流系統的運動場結構與微物理過程有一定差異，而這些差異有利於有 System edge Cls 的系統增強。

雙偏極雷達觀測結果顯示有較多 System edge Cls 系統有較強並水平延伸之後方入流（RIJ）及較明顯的軟雹/冰雹淞化成長過程（Riming growth），顯示（二）可能扮演較重要的角色。我們可以利用 WRF 微物理敏感度實驗對假設（二）作進一步的討論。實驗結果顯示，當實驗容許軟雹或冰雹可沿大小不同以不一終端速度下降的時候，後方入流會有增強並於水平方向能有所延伸。同時，RIJ 的水平延伸容許更多對流於系統邊緣產生。

上述結果出現的原因與對流系統內熱力場的變化有關：密度較大（較重）的軟雹粒子較大的終端落速使這些粒子能於較靠近對流上升區（Updraft region）的位置降到 0℃線以下。軟雹溶解過程中的潛熱消耗可增大對流上升區溶解層上下的浮力差並產生較明顯的負浮力氣壓擾動（negative buoyancy pressure perturbation），此相對低壓對 RIJ 增強有主導影響。實驗結果同時顯示較強的 RIJ 可將部分下降至 0℃ 線附近的軟雹重新帶到上升氣流中並可以增強淞化成長，而令更多軟雹能在上升區中形成。這過程中所產生額外的潛熱釋放可使對流系統強度增加並有更大維持能力。

Keywords: 降水過程; 雲微物理; 中尺度動力機制

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Chapter 1 – Introduction and Motivation

1.1. Nocturnal convective systems over the Great Plains

The spatial and diurnal characteristics for warm season precipitation in the continental U.S. varied within the continental U.S during boreal summers (JJA). Late afternoon precipitation dominates in both the southeastern U.S. and above the Rocky Mountains while precipitation in the Great Plains tends to occur near midnight (Fig.1.1.1). The nocturnal peak could be attributed to the upscale growth and propagation of eastward-propagating mesoscale convective systems (Fig.1.1.3). Compared to daytime thunderstorms, the predictability of these nocturnal systems remained low (Davis et al. 2003) due to insufficient understandings on their responses to the ambient environment and their internal dynamical and microphysical structures.

1.2. Current understandings on nocturnal convective systems

The most crucial difference between daytime convective systems and their nocturnal counterparts lies in the origin of the system instability and the lifting potential for convective cold pools. Daytime convective systems are primarily sustained by instability within the boundary layer (“surface-based”) while nocturnal systems are at least partially sustained by instability above the boundary layer (“elevated”).

Whether a convective system is surface-based or elevated is of substantial importance since the propagation speed for surface-based systems could be reasonably predicted by cold pool dynamics (Bryan et al. 2005); these systems could also cause severe near-surface wind. For
elevated systems, however, their propagation speed often departure from values predicted by cold pool dynamics (French and Parker 2010). These systems could also produce weak near-surface wind due to the inability for convective downdrafts to penetrate to the surface.

While surface-based systems generally transition to elevated systems as nocturnal boundary layers strengthen, the specific transition time often varies from case to case. Indeed, a series of idealized simulations (Parker 2008; French and Parker 2010) indicated that nocturnal systems could remain partially surface-based as long as the boundary layer was not completely devoid of CAPE (Convective Available Potential Energy). In terms of system propagation speed, their simulated system decelerated during the initial boundary layer cooling while accelerated afterwards. The initial deceleration could be attributed to the relaxation of density gradient across cold pool while the acceleration occurred since the system was sustained by wave-like structures during this period.

Whether a convective system is completely elevated or partially surface-based is also of importance due to their different precipitation potential. A recent modeling study by Schumacher (2015) indicated that elevated systems featured less accumulated precipitation compared to surface-based systems.

1.3. Motivation for this study

While the above studies revealed critical characteristics for nocturnal convective systems, substantial uncertainties remained. Specifically, there had not been enough studies on the role of the kinematic and microphysical structures of these systems. We also lack understanding on how much the internal processes impact the evolution of these systems.
To alleviate this, the environmental evolution and the internal kinematic and microphysical evolution of the PECAN 15 July 2015 MCSs were analyzed. The same case was utilized in Grasmick et al. (2018; *G18*) to illustrate correlations between MCS convective structure and local cold pool variabilities. Our study expanded upon their study in discussing some aspects left unexplored in their observational analysis:

- The research focus of *G18* was on the interaction between local variations in cold pool strength and the relative CI locations in the convective system. The origin of such cold pool variations (which was largely induced by microphysical processes), was left unexplored in their work. Our current work also expands upon *G18* in attempting to understand why system rear CI weakened at a quicker rate than system edge CI. Specific focus will be paid to the ice microphysical-kinematic interaction within the systems.

- *Mesoscale moisture transport* associated with the developing low-level jet (LLJ) and its role in regulating the environmental instability profiles was not discussed in Grasmick et al. (2018), which mainly focused on its dynamical contribution in bore generation.
Chapter 2 - Event Overview

1. Synoptic Overview

The RAP analysis with horizontal resolution of 13km show that the synoptic environment is broadly conductive to widespread convective development in the Great Plains. A broad mid- and upper-level trough located off the coast of California at 1200UTC on 14 July 2015 (Fig. 2.1.1a,b) provided positive vorticity advection in the Plains. The synoptic flow also transported deep tropic moisture from Pacific and the developing Hurricane Dolores (2015) into the region, which was confirmed by satellite water vapor imagery (Fig.2.1.2a). Increased moisture availability in the western US undoubtedly played a role in initiating convection over the elevated terrain in CO. Elsewhere, the southerly wind branch associated with the deep anticyclonic circulation in Texas can locally enhance the LLJ in eastern CO, western Kansas, and northern Oklahoma after sunset (Fig. 2.1.1c).

2. Radar morphology and evolution of observed nocturnal systems

The IOP30 systems could be traced back to two separate convective clusters in north-eastern and south-eastern Colorado that grew upscale in environments with deep and well-mixed CBL (convective boundary layer). These systems acquired different morphology by sunset due to different deep-layer shear strength locally (Fig.2.2.2), with the south-eastern MCS (MCS$_{S1}$) acquired supercellular characteristics$^1$ and the north-eastern MCS (MCS$_N$), quasi-linear (Fig.2.2.3a). By 0255UTC, a radar fine line formed due to a rearward shift in convective initiation (CI) locations in MCS$_N$ (Fig.2.2.3c). This radar fine line represented the

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$^1$ Rotating updraft was observed in NEXRAD 0.5° scan of radial velocity from Dodge City, Kansas (KDDC). (not shown)

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cold pool edge since there was an approximately 5°C temperature drop across this feature. MCS\textsubscript{N} started weakening after this rearward shift in CI location and most convection deteriorated into stratiform by 0355 UTC.

The MCS\textsubscript{S1}, on the other hand, featured CIs located near the gust front. In contrast to the MCS\textsubscript{N}, MCS\textsubscript{S1} showed no sign of weakening, even forming a bowing segment by 0430 UTC (Fig.2.2.3e). The cold pool is locally stronger here with stronger near-surface wind (>16 m/s 5-min averaged wind at Ellis, Kansas) owing to the descending rear-inflow jet (RIJ). For the MCS\textsubscript{S2}, the weaker cold pool for this system suggested sustenance partially through bore lifting (Grasmick et al. 2018). MCS\textsubscript{S1} and MCS\textsubscript{S2} merged to form a large, bowing MCS shortly after 0720UTC (Fig.2.2.3h).

3. Two hypotheses on the different system evolutions at night

A crucial issue that we will be focusing on this case study is the difference in physical processes that contributed to the weakening of MCS\textsubscript{N} CIs and the strengthening of MCS\textsubscript{S} CIs during the nocturnal hours. Here, we propose some possible hypotheses that might cause the different system evolution during IOP30:

1. **Mesoscale variabilities in moisture and temperature above boundary layer near the two systems.**
2. **Potential contributions of mixed-phase microphysical processes.**
Chapter 3 – Instrumentation and Experimental Design

3.1. Instrumentation platforms and observational data quality control

During the PECAN field campaign, multiple observational platforms were deployed to provide a rich dataset on nocturnal MCS characteristics. The locations of the available observation platforms during IOP30 are provided in Fig.3.1.1.

1. Summary of utilized platforms

(a) Radiosondes

High temporal-resolution radiosonde datasets from the western half of the PECAN domain were used to provide information on the transient changes in the pre-MCS environments. Data collected by two different PECAN integrated sounding arrays (PISAs) at FP3 and FP5 were analyzed for this study. We also augmented our dataset by adding soundings from mobile PISAs near Scott City, KS (MP3; Wagner et al. 2016b). All of these sounding platforms utilized Vaisala RS92 rawinsondes. Data gaps associated with on-site issues were rectified by a simple time-height interpolation process.

The temporal scale of the radiosonde dataset obtained during IOP30 is generally on the order of 40-60 min, which permits us to examine the pre-MCS environmental evolution in a relatively detailed manner.

(b) Lidars

Low-tropospheric moisture evolution could be observed in a spatially and temporally
detailed manner with lidars. Lidars used in this study included: micro-pulse lidars (MPLs) located at FP3 and CL31 ceilometers at FP4. One caveat in utilizing the ceilometer and MPL datasets is that these instruments cannot directly evaluate environmental variables (e.g., temperature, moisture). We are not aware of algorithms that link the lidar observed variables (e.g. attenuated backscatter coefficient) and environmental variables.

(c) Radars

The radars used in this study included the two operational NEXRAD radar at Goodland, KS (KGLD) and Dodge City, KS (KDDC), the ground-based research S- and K-band radar (SPolKa) south of FP3. All radars possess dual-polarization capabilities.

The radar scanning strategy adopted by the NEXRAD radars during PECAN was volume coverage pattern (VCP) 12. VCP 12 takes ~4.5 minutes to complete and consists of surveillance scans at 14 different elevated angles from 0.5° to 19.5°. The complete scanning procedure for the SPolKa comprises two modes: PPI (Plan Position Indicator) and RHI (Range Height Indicator). RHI scans are available every 30 degrees.

3.2. Basic WRF settings

1. WRF specifications

In this study, Version 3.9 of WRF-ARW model (Skamarock et al. 2008) was used to simulate the two nocturnal MCSs described in Chapter 2. The dynamics core of this model are
compressible\textsuperscript{2}, non-hydrostatic Euler equations. The numerical methods used to solve these differential equations are the third-order Runge-Kutta schemes. A time-split approach was used to circumvent the time step limitation by high-frequency acoustic modes.

The horizontal spatial differentiation terms in the governing equations could be estimated by utilizing the Arakawa C grid staggering under different map projection systems\textsuperscript{3}. Under C grid staggering, variables related to velocity \((u,v,w)\) are separated from thermodynamic variables by 1/2 grid length, as shown in Fig.3.2.1. The points where thermodynamic variables are located are denoted as \textit{mass points}, and these points are also the locations at which moisture variables are defined and diagnostic variables (e.g. pressure, inverse density) are calculated (Skamarock et al. 2008).

2. Domain design and boundary conditions

The domain setting applied in this study is a triple-nested domain (Fig.3.2.2; Table 3.2.1) with an outer domain with 27-km horizontal grid spacing and two inner domains with 9-km and 3-km horizontal grid resolution, respectively. Two-way feedback is enabled for the two inner domains. While a grid size on the order of 100-m is needed to directly simulate individual convective cells (Bryan et al. 2003), a 3-km grid spacing could reasonably represent the overall MCS structure. All domains contain 42 vertical levels with model top located at 10hPa. 0.75\degree x 0.75\degree EC-Interim gridded reanalysis data was used as the initial and lateral boundary conditions (ICs and LBCs). The simulations were initialized at 1200UTC.

\textsuperscript{2} The use of compressible governing equations indicates that sound waves exist in the WRF physical solutions.\textsuperscript{3} In WRFV3, four types of projection systems are supported: Lambert conformal, polar stereographic, Mercator, and latitude-longitude projections.
on 13 July 2015 and ran continuously for 48 hours. A relatively long leading time was utilized in our study to account for model spin-up.

3. Physical parameterization options

The model physical parameterization schemes utilized in this study are: Dudhia shortwave radiation scheme, RRTM longwave radiation scheme, Yonsei University planetary boundary layer (PBL) scheme (YSU), MM5 Similarity Scheme and Unified Noah Land Surface Model, and Kain-Fritsch cumulus parameterization scheme\(^4\), respectively.

For nocturnal conditions, local-closure PBL schemes (e.g. MYJ) are somewhat preferable since they are better in capturing the structure of NBLs (Shin and Hong 2011). Despite this, YSU scheme was used in our simulations since the stabilization of pre-MCS\(_3\) boundary layer occurred very slowly during IOP30 and convective boundary layer characteristics remained during the early evening period.

Since the purpose of this study was not to provide a comprehensive comparison on the relative performance for different schemes, only two microphysical schemes were tested (MORR and NSSL) in this study and the NSSL scheme was found to overperform the MORR for this specific event. We do note that it is difficult to confirm that our combination of physical schemes is optimal without robust comparisons. Nevertheless, we do argue that our current physics combination is reasonable and adequate for our research purposes.

\(^4\) Used only in the outermost domain.

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Chapter 4 – Observational Analysis

4.1. High-frequency Radiosonde Analysis

1. Pre-MCS environmental comparison

The differences between the two pre-MCS environments could be summarized by their respective $\theta_e$ profiles prior to system arrival\(^5\). For the pre-MCS\(_N\) environment (FP5; Fig.4.1.1a), $\theta_e$ profile from 2359 to 0200 UTC showed a $\sim$0.8 km descent of a mid-level inversion layer, suggesting that the pre-MCS\(_N\) environment was primarily impacted by subsidence motion. For the pre-MCS\(_S\) environment (MP2; Fig.4.1.1b), no inversion layers could be identified and the boundary layer top height increased with time, both illustrated a lack of subsidence motion. These differences caused considerable discrepancies in the moisture characteristics above boundary layer: with the pre-MCS\(_S\) environment moistened above boundary layer while the pre-MCS\(_N\) environment became drier prior to gust front passage.

The above discussions indicated that mesoscale environmental heterogeneities in mesoscale motions likely played a role in creating environmental variabilities at night. Sustained subsidence stabilized the pre-MCS\(_N\) environment (Section 4.1.3) through inversion layer intensification and moisture depletion. In contrast, the pre-MCS\(_S\) environment destabilized due to the moistening by sustained upward motion.

\(^5\) Relative locations between the sounding launch site and the convective systems are summarized in Appendix 2.
2. Quasi-two-dimensional analysis

Sounding data were interpolated onto a constant time-height grid in this section to examine the transient changes of atmospheric variables with time. Pre-MCSN environmental potential temperature perturbation evolution (Fig.4.1.4a) shows the aforementioned low-level subsidence generated a ~3-5°C warming between ~2-4 km from 0000 to 0230 UTC while the boundary layer cooled gradually. The sustained subsidence also depleted the environmental moisture considerably (Fig.4.1.4b). In contrast, upward motion cooled (Fig.4.1.5a) and moistened (Fig.4.1.5b) the pre-MCSs environment above boundary layer.

Both environments experienced further cooling after the arrival of the gust front. The comparison of on-site surface observations indicate that the surface temperature drop associated with MCSs cold pool was generally stronger than the MCSN cold pool (Fig.4.1.3). Such differences could likely be attributed to internal microphysical processes rather than ambient environmental properties since the near-surface RH were similar between the two environments by ~0300UTC (Fig.4.1.2).

3. Comparison of pre-MCS environmental instability parameter profiles

In this section, CAPE, CIN and ΔZLFC vertical profiles will be used to access the impacts of the aforementioned environmental variabilities on the environmental instabilities. ΔZLFC represents the distance between the originating height and LFC of a given parcel. This quantity was used in Peters et al. (2017) as a proxy for the lifting needed to initiate convection and would indicate unstable parcels if parcel lifting (Δz) exceeds ΔZLFC.

Despite the sustained subsidence, pre-MCSN CAPE (Fig.4.1.6a) only decreased slightly in
from 2359 to 0230 UTC. On the other hand, CIN nearly doubled for the pre-MCSN environment after 2359 UTC (Fig.4.1.6b), which hindered the ability for MCSN to release instability. This was represented by a ~1-2 km $\Delta Z_{LFC}$ increase below 2.5 km AGL (Fig.4.1.6c). Assuming parcels mainly underwent adiabatic ascent in the pre-convective environment, the apparent cold pool lifting could be roughly estimated with the isentrope displacement in Fig.4.1.4 to be ~1 km. This lifting was insufficient for boundary layer parcels and for most elevated parcels apart from a narrow layer between 1.75 and 2.25 km AGL. The rather small CAPE within this layer could potentially explain the continued MCSN weakening during nocturnal hours.

Contrastingly, the pre-MCSs CAPE (CIN) steadily increased (decreased) with time (Fig.4.1.7). We also note that the CAPE values within the layer where $\Delta Z_{LFC} < \Delta z$ were substantially larger (~1500 J kg$^{-1}$) than for the pre-MCSN environment. The dramatically higher elevated CAPE values in the pre-MCSs environment was likely conducive to the initial development of MCSs.

### 4.2. Pre-convective environmental characteristics – Potential contributors for environmental heterogeneity

#### 1. Inference of mesoscale vertical motions

Enhanced moisture layer depth in the pre-MCSs environment and the formation of the inversion layer in the pre-MCSN environment hinted at the potential role of mesoscale vertical motions in inducing the observed environmental variability. In this section, we will provide a diagnostic study on the physical characteristics of these vertical motions.
(a) Vertical motion retrieval method

Vertical profiles of pressure velocity were computed kinematically following the methodology outlined by Trier et al. (2017; T17 in the remainder of this section), which is based on Bellamy (1949; B49 in the remainder of this section). Pressure velocities at any given isobaric surfaces by integrating the continuity equation,

$$\omega(p_2) = \omega(p_1) - \int_{p_1}^{p_2} v_p \cdot \vec{V} \, dP \quad (Eq. 4.2.1),$$

where $p_2$ is the pressure value at the tropopause and $p_1$ is the pressure value at the isobaric surface closest to surface. $v_p \cdot \vec{V}$ could be estimated by the instantaneous area change rate for the triangle bounded by three radiosonde launch sites,

$$v_p \cdot \vec{V} = \frac{1}{A} \frac{DA}{Dt} \quad (Eq. 4.2.2),$$

The vertical air pressure and horizontal wind profiles for the three sites were first interpolated to a common pressure coordinate bounded at 250 and 950 hPa, and with a 5 hPa vertical resolution. The instantaneous triangle area change rate could be estimated by the area difference between the original triangle and the area of the triangle after being perturbed by the observed wind over an arbitrary time period ($\Delta t$).

(b) Analysis result and discussion

(i) Pre-MCS environments

Figure 4.2.1 presents both the raw and corrected horizontal divergence vertical profiles valid

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6 The diagnosed $v_p \cdot \vec{V}$ is insensitive to the length of $\Delta t$ assuming the same $\Delta t$ was used during the perturbation process.
7 The resulting $\nabla \mu \vec{U}$ after subtracting a portion of the raw $\nabla \mu \vec{U}$ field following the correction scheme detailed
at the T1 and T2 triangle centroids (location denoted in Fig.4.2.2a), which approximately represented the pre-MCSN and pre-MCSs environments, respectively. Above the T1 centroid (Fig.4.2.1a), there was $\nabla_h \bar{U}$ convergence (divergence) above (below) ~800 hPa and downward motion in the low troposphere. Above the T2 centroid (Fig.4.2.1b), the diagnosed $\omega$ was weakly negative in the lowest 200 hPa, which was consistent with a continuously destabilizing low troposphere. The diagnosed $\omega$ profiles also demonstrated stronger upward motion above ~650 hPa. However, caution is needed when interpreting the diagnosed vertical motion aloft since the accuracy of the diagnosed vertical motion could be impacted by the considerable time difference\(^8\) between the utilized soundings.

(ii) Ambient environment

In this section, the $\omega$ profiles within triangles T3 and T4 (locations denoted in Fig.4.2.2b) are compared. The orientation of these two triangles approximately paralleled the 700-850 hPa LLJ direction (Fig.4.2.2b). The diagnosed $\omega$ vertical profiles at T3 and T4 centroids (Fig.4.2.3) showed a meridional gradient in updraft motion strength aloft: the T3 profile exhibited a core of upward motion of $\sim - 3.5 \mu b s^{-1}$ above 800 hPa while the vertical motion above T4 centroid was weak below 700 hPa. The impact of this apparent LLJ-paralleled mesoscale motion variability on the temperature and moisture distributions in west-central Kansas will be discussed with gridded RAP analysis in the next section.

2. Generation of environmental heterogeneity: Observational Analysis

(a) RAP analysis

\(^8\) The SPARC sounding launch closest to 0400 UTC was not used in our calculation since it was launched shortly after to the MCS\(_s\) passage time. 0300 UTC SPARC sounding was used instead.
Our first discussion in this section involves the surface moisture and temperature distributions (Fig.4.2.4) in west-central Kansas. No signs of near-surface baroclinicity existed apart from a weak quasi-stationary frontal area in southwestern NE, which was too far north. On the other hand, the near-surface moisture distribution showed a meridional gradient, with a prominent SE-NW moist corridor in west-central KS.

The collocation between 700-850 hPa warm air core (Fig.4.2.5a) and the southerly LLJ in the OK Panhandle generated substantial warm advection, which subsequently created upward motion in the pre-MCS environment (Fig.4.2.5c).

The overlaps between amplified moisture variability (~15mm precipitable water difference between pre-MCS environments; Fig.4.2.5d) and upward motion (Fig.4.2.5c) pointed to the crucialness of vertical moisture advection for our case. We evaluate the above statement by considering the water vapor mixing ratio change caused purely by advection.

\[ \frac{\partial q_v}{\partial t_{adv}} = -\mathbf{V} \cdot \nabla p q_v - \omega \frac{\partial q_v}{\partial p} \quad (Eq. 4.2.5) \]

The first term on the left-hand-side of (Eq.4.2.5) represents the horizontal moisture advection while the second term represents the vertical moisture advection. Comparison between the moisture distributions (Fig.4.2.5b,d) and the advection terms (Fig.4.2.6) at 0400 UTC suggests that horizontal moisture advection could not locally moisten the pre-MCS environment due to the substantial poleward transport. In contrast, vertical moisture advection was negative (upward) within the pre-MCS environment, which enabled moisture accumulation.

The vertical motion variability was intrinsically related to the mesoscale kinematic structure.
The overall flow pattern above boundary layer was defined by horizontal deformation in the region, with the axis of contraction approximately paralleled the horizontal temperature gradient. These flow characteristics could induce a frontogenetic transverse circulation. To confirm this, low-level frontogenesis was calculated from the gridded RAP data (Fig. 4.2.7a). The frontogenesis bands were paralleled to (perpendicular to) the vertical motions bands (low-level temperature gradient). These bands (Fig. 4.2.7b) were limited to 700-850 hPa, with signs of a transverse circulation\(^9\) formed across the bands.

(b) **PECAN observations**

Pre-MCS\(_S\) environment was shown, in the above analyses, to be destabilized by sustained upward motion initiated within the LLJ, locally enhanced along the updraft branch of a traverse circulation. These findings could be corroborated by PECAN lidar observations. Figure 4.2.8a shows the attenuated backscatter coefficient (\(\beta\)) observed by the CL31 ceilometer at FP4. Height evolution of the lidar-observed high backscatter layers could be utilized to indirectly infer the existence of weak vertical motions (Shapiro et al. 2018). Two conclusions could be drawn from this figure: (a) Environment downstream of the elevated frontogenesis was highly stratified, and (b) Upward motion was weak in this region. For the pre-MCS\(_S\) environment (FP3), however, vertically-uniform aerosol distribution and routine aerosol overshooting both illustrated a more convectively-active boundary layer (Fig. 4.2.8b) upstream of the elevated frontogenesis during the early evening period. The aerosol layer also deepened with time, with was consistent with the RAP analysis.

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\(^9\) Upward (downward) motion in the southwest (northeast) half of the cross-section.
4.3. Internal microphysical and kinematic characteristics

In this section, we will examine the differences in storm microphysical and kinematic characteristics between the two convective systems when they developed under contrasting ambient environments. Such differences are likely critical in (a) creating the system cold-pool variability, and (b) maintaining MCSs when it moved to drier environment in central Kansas. The dual-polarized radar variables that we will be analyzing in the remainder of this section include $Z_{HH}$ (Horizontal reflectivity), $Z_{DR}$ (Differential reflectivity), $K_{DP}$ (Specific differential phase), $\rho_{HV}$ (Correlation Coefficient).

1. Reflectivity and kinematic characteristics of the analyzed systems

(a) MCS$_N$

Figure 4.3.1 shows the reflectivity and radial velocity fields observed by KGLD at 0300 UTC, which was shortly before the MCS$_N$ weakening phase. In terms of its kinematic characteristics, the comparison between 0.5$^\circ$ (Fig.4.3.1b) and 1.3$^\circ$ (Fig.4.3.1c) velocities suggest the cold pool height was less than ~1 km AGL$^{10}$ north of KGLD while slightly deeper west of the KGLD. Vertical cross sections along the 270$^\circ$ azimuth angle (Fig.4.3.2) indicate slightly deeper rear inflow behind the system due to the southwesterly synoptic flow. The rear inflow within MCS$_N$ was mostly confined to the lowest 1 km and was relatively weak near the melting level, however.

$^{10}$ Since the velocity direction shifted from directed towards radar at 0.5$^\circ$ to directed out of the radar at 1.3$^\circ$ near 39.6 N (~30 km north of the radar).
(b) MCSs

Figure 4.3.3 shows the horizontal reflectivity and radial velocity structures of mature phase MCSs observed by the SPolKa radar near FP3 at 0600 UTC. The mature MCSs was characterized by the development of a bowing segment near the cold pool edge and formation of a transition zone. Vertical cross section of radial velocity along the 300° azimuth (Fig.4.3.4b) shows near-surface rear-to-front flow of >30 m s⁻¹ (~10 m s⁻¹ larger than MCSN) behind the cold pool edge, which implied stronger cold pool. Apart from the near-surface RTF flow, RTF inflow near melting level was substantially stronger for MCSs. Figure 4.3.4b also shows greater lifting when the inflow layer encountered the stronger elevated rear inflow.

2. Microphysical characteristics of the analyzed systems

(a) Horizontal structure of dual-polarization variables

Figure 4.3.5 (Figure 4.3.6) shows the PPIs for \(Z_{DR}\), \(\rho_{HV}\), and \(K_{DP}\) observed from KGLD (SPolKa) at 0.5° angle, respectively. The \(Z_{DR}\) of the system rear CIs were generally between 1.5 and 2 dB. In contrast, the \(Z_{DR}\) in the edge of MCSs exceeded 3 dB, which indicated larger raindrops near surface. In terms of hydrometeor type, the relatively weak \(Z_H\) (Fig.4.3.1a) and the \(\rho_{HV}\) uniformity (Fig.4.3.5b) for MCSN suggest that the system was dominated by smaller rainwater particles near surface. For MCSs, however, the \(\rho_{HV}\) for the cells along MCSs leading edge was generally smaller than one. The lowered \(\rho_{HV}\), enhanced \(Z_H\) and \(Z_{DR}\) point to the co-existence of heavier melting rimed particles and large raindrops near surface. Finally, the comparison of \(K_{DP}\) also demonstrates lower rainwater concentration for MCSN than MCSs.

To conclude, the dominant hydrometeor for the weakening MCSN was smaller raindrops while MCSs featured (a) melting rimed particles and (b) large raindrops near surface.
(b) *Vertical profiles for dual-polarization variables: $Z_H$ and $Z_{DR}$*

Dual-polarized variables are first interpolated onto Cartesian coordinates with a horizontal grid spacing of 500 m and a vertical grid spacing of 250 m over a 13.5kmx300kmx300km volume with the Python Atmospheric Radiation Measurement (ARM) Radar Toolkit (PyART) module, which was then used to derive the averaged vertical profiles for the two systems (the domains used were shown in Fig.4.3.7).

$Z_H$ profiles within the analysis domains (Fig.4.3.8) were largely similar, with both lacking $Z_H$ maxima near the melting level. Nevertheless, two differences could be identified, including:

- Sharper $Z_H$ drop with height between ~3-5 km AGL for MCSN.
- MCSN $Z_H$ increased towards ground in the lowest 2 km AGL while slightly decreased for MCSs.

The microphysical processes that caused these differences could be determined by examining the dual-polarization variable profiles. The average $Z_{DR}$ (Fig.4.3.10a) slowly increased towards ground for MCSN below melting level while the $Z_{DR}$ increase tendency was stronger for MCSs. The slow $Z_{DR}$ and $Z_H$ increase for MCSN reflected the collision-coalescence growth of raindrops below the melting level (Kumjian and Prat 2014). On the other hand, the combination of $Z_H$ decrease and $Z_{DR}$ increase for MCSs was inconsistent with coalescence growth and suggested that other microphysical processes might be at work. Since $Z_H$ is sensitive to concentration and size while $Z_{DR}$ is sensitive only to size, the observed $Z_H$-$Z_{DR}$ structure indicated the MCSs was mainly populated by a small concentration of oblate (i.e. larger) raindrops near surface. The dominance of larger particle near surface could be

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attributed to size sorting, which caused heavier raindrops to fall at a quicker rate towards the ground. Elsewhere, $Z_{DR}$ enhancement near melting level reflected the refractivity increase for melting particles (Kumjian 2013). The $Z_{DR}$ increase for MCSS occurred at a more vertically-expanded layer compared to MCSN. The broader $Z_{DR}$ increase layer signified heavier ice particles for MCSS due to their higher fall speed and could melt over a deeper vertical layer.

The differences in $Z_{DR}$ were indicative of distinct ice/mixed phase processes aloft for the two systems. For MCSS, the layer with enhanced $Z_{DR}$ extended from the melting level to ~6 km AGL, which encompassed the temperature range of secondary production of columnar ice crystal during riming process (Sinclair et al. 2016). We could infer that mixed phase processes were robust in the MCSS convective region. The prevalence of mixed-phase growth also implies stronger updrafts for MCSS, where the columnar ice crystals can grow rapidly into large rimed particles within supercooled drops-rich updrafts, decreasing the $Z_{DR}$ values\textsuperscript{11} above ~6 km AGL. In contrast, the larger $Z_{DR}$ in MCSN above 6 km AGL implies greater ice crystal anisotropy (Didlake and Kumjian 2017). Crystal habit diagram (Fig.4.3.10) indicated dominant plate-like crystal with dendritic growth in this temperature range, which enhanced the $Z_{DR}$.

\textit{(c) Vertical profiles for dual-polarization variables: $\rho_{HV}$ and $K_{DP}$}

The $\rho_{HV}$ profiles (Fig.4.3.9b) consistently showed values close to unity for MCSN, which implied size and type uniformity for MCSN hydrometeor aloft. The ice particles formed in MCSN were of lower densities, as indicated by a sharp and narrow $\rho_{HV}$ minimum below ~4

\textsuperscript{11} Existence of rimed particles within MCSS updrafts was supported by increased LDR (Linear Depolarization Ratio) above ~6 km AGL. These results were not shown since LDR was not available for KGLD.
For MCSS, the steady T drop towards surface indicated the presence of melting rimed particles below the melting level. Additionally, the enhanced vertical extension of small T values above melting level was in line with greater concentrations of supercooled drop and rimed particle within the MCSS convective updrafts. Finally, the KDP profiles (Fig. 4.3.9c) exhibited dramatically higher KDP for MCSS below the melting level, which suggested a boost in rainwater concentration and liquid water content due to the microphysical changes aloft. KDP above melting level is a suitable tool in analyzing these changes since it is sensitive to the concentration of secondary ice particles. The average KDP value for MCSs was ~0.2° km^{-1} between -3° and -8°C, which was similar to the observed KDP values in secondary ice production regions (Sinclair et al. 2016). KDP decreased to ~0.05° km^{-1} above 6 km AGL due to the riming growth of secondary ice crystal and loss of non-sphericity. For MCSs, the generation of ice crystals with high anisotropy enhanced both the ZDR and KDP above 6 km AGL due to the riming growth of secondary ice crystal. The KDP profile was slightly less vertically extended compared to the ZDR profile. Specifically, comparison on the KDP and ZDR vertical gradients (not shown) suggests an approximately 1 km lowering for the KDP compared to the ZDR. Since KDP is sensitive to hydrometeor concentration, the height difference illustrates the gradual concentration increase for plate-like crystals towards the melting level, whereas the higher ZDR profile reflected the oblateness of the crystal formed near -15°C. Finally, the KDP below the melting level was much smaller for MCSs than MCSs due to smaller rainwater concentrations in the lowest 0.5 to 1 km AGL, the KDP reduced slightly towards ground for the MCSs while it enhanced slightly for MCSs taking the above discussion into account, the reduction of KDP for MCSs was consistent with the size sorting of raindrops and the depletion of small drops by evaporation. The upick in KDP for MCSs likely result from (a) weaker rain and rimed particles below the melting level. Additionally, the enhanced vertical extension of small T values above melting level was in line with greater concentrations of supercooled
drop evaporation despite the smaller drops in the system, and (b) limited size sorting for rain drops.

3. Summary

Dual-polarized variable profiles indicate that cold pool was stronger for MCS\textsubscript{S} through rain evaporation and rimed particle melting within the lowest 0.5-1 km. The dominant microphysical process near surface for MCS\textsubscript{N}, on the other hand, was likely weak raindrop collision-coalescence.

The parameter profiles aloft, on the other hand, pointed to a more robust mixed phase growth within MCS\textsubscript{S}. The higher $Z_{\text{DR}}$ and $K_{\text{DP}}$ between -3 and -8\degree C illustrated an amplification of secondary ice production in MCS\textsubscript{S}, which could grow rapidly into heavier graupel or hail particles. These mixed phase processes primarily occurred near the top of the $Z_{\text{DR}}$ and $K_{\text{DP}}$ columns near the system edge (Fig.4.3.11). For MCS\textsubscript{N}, higher $Z_{\text{DR}}$ and $K_{\text{DP}}$ values occurred in the planar ice crystal formation layer at \sim-15\degree C. These ice particles in MCS\textsubscript{N} are lighter and descended at a slower rate than the rimed particles in MCS\textsubscript{S}. The profiles also indicated a potential link between the concentration of mixed phase (riming) particles aloft and the rainwater concentrations below the melting level.
Chapter 5 - Evolution of simulated MCS and moisture variability

5.1. Description of simulated storm evolution

In the FULL experiment, two convective clusters developed in southeastern and northeastern CO at ~0100 UTC (Fig. 5.1.1a) and swiftly grew upscale by ~0200 UTC (Fig. 5.1.1b). The simulated MCS\textsubscript{S} structure compared favorably with observations. Similarly, the weakening tendency for MCS\textsubscript{N} was also well captured by WRF. The strengthening phase for simulated MCS\textsubscript{S} culminated in the formation of a bowing segment at ~0400 UTC (Fig. 5.1.1c).

Encouragingly, MCS\textsubscript{S} maintenance after the early evening period was well predicted by the WRF model, as shown by the high reflectivity values at ~0500 UTC (Fig. 5.1.1d). Similar to the observed system, there was a tendency for strongest convection to aggregate in the fourth quadrant within the propagating MCS\textsubscript{S}\textsuperscript{12}. Despite relatively similar system structure, we note a consistent westward location bias for our simulations, which reflected an underestimation in environmental instability. Despite these errors, we argue that the usage of these simulations is acceptable since the primary objective for these simulations is to compare systems that are otherwise similar apart from rimed particle characteristics.

5.2. Simulated environmental heterogeneity and moisture transport

Observational analyses indicated that environmental variabilities was mostly induced by

\textsuperscript{12} Assuming the centroid of MCS\textsubscript{S} is the origin of a Cartesian coordinate system.
moisture above the boundary layer. The criticalness of moisture could be confirmed by the
~1.5 g kg\(^{-1}\) higher water vapor mixing ratio in the pre-MCS\(_S\) environment (Fig.5.2.1b) and
the highly similar pre-MCS \(\theta\) profiles (Fig.5.2.1a) above ~1 km AGL.

In this section, various moisture-related variables will be used to show that, similar to the
observations, vertical moisture transport was critical in generating the moisture variabilities
in western KS for the WRF simulations.

(a) Integrated Vapor Transport (IVT) and Integrated Water Vapor (IWV)
The spatial characteristics and strength of poleward moisture transport can be quantified by
examining the column-integrated vapor transport (IVT) as follows:

\[
IVT = \sqrt{\left(\frac{1}{g} \int_{p_{sfc}}^{P_{top}} q_v u \, dp\right)^2 + \left(\frac{1}{g} \int_{p_{sfc}}^{P_{top}} q_v v \, dp\right)^2} \quad (Eq. 4.2.1)
\]

where \(q_v\) is the model-predicted vapor mixing ratio, \(u\) (\(v\)) is the zonal (meridional) wind
component, \(g\) is the gravitational acceleration and \(dp\) is the pressure difference between two
adjacent model levels. Environmental moisture availability can also be quantified by column-
integrated water vapor (IWV) as follows:

\[
IWV = \frac{1}{g} \int_{p_{sfc}}^{P_{top}} q_v \, dp \quad (Eq. 4.2.2)
\]

At 0100 UTC (Fig.5.2.2a), IVT was generally weak and exhibited a higher spatial correlation
with the surface wind (Fig.5.2.2b). By 0400 UTC, the IVT strengthened to ~425 kg m\(^{-1}\)s\(^{-1}\) in
west-central KS but weakened over southwestern KS (Fig.5.2.2c). This change was related
to the LLJ since (a) the IVT distribution was more closely correlated to LLJ (Fig.5.2.2d) over
central KS and (b) LLJ strengthened and expanded considerably from 0100 to 0400 UTC.

Interestingly, the 0400 UTC IWV (Fig.5.2.3b) was negatively correlated (IVT high -> IWV low, and vice versa) with IVT (Fig.5.2.2c). This negative correlation attested to the limited importance of horizontal moisture transport in local environmental moistening process. Overall, the simulated IVT and IWV patterns are similar to the RAP analyses (Fig.5.2.4a,b).

**b) Moisture convergence**

The negative correlation between IWV and IVT indicates that deeper moisture layers might be more important than horizontal moisture transport in destabilizing the pre-MCS environment. Insights on regional moisture variability could be obtained from examining the mesoscale moisture convergence component in the environmental moisture budget equation (Trenberth et al. 2011),

\[
\frac{\partial w}{\partial t} + \nabla \cdot \left( \frac{1}{g} \int_0^{P_{sfc}} vqdP \right) = E - P \quad (Eq. 4.2.3)
\]

where the terms on the left-hand side of Eq.(4.1.3) represent (i) time evolution of precipitable water and (ii) moisture divergence/convergence, respectively. The terms on the right-hand side of Eq.4.1.3 represent surface evaporation \((E)\) and surface precipitation rate \((P)\).

At 0100 UTC, west-central KS was dominated by moisture divergence (Fig.5.2.5a), which adversely impacted moisture accumulation over the region. By 0400 UTC, however, moisture convergence dominated over much of central KS (Fig.5.2.5b). It is obvious from Fig.5.2.2-16 that the regional distribution of moisture convergence/divergence was controlled by the spatial distribution of low-level wind speed. Regions with stronger low-level wind maxima
were characterized by higher IVT, slightly lower IWV and negative moisture convergence (moisture divergence); such regions are less conducive to environmental destabilization due to robust horizontal moisture transport out of the atmospheric column.

(c) Vertical motion

The 0400 UTC pressure vertical velocity ($\omega$) at 750hPa showed weak upward motion (<1 Pa/s) over much of west-central KS (Fig. 5.2.6a). These weak updrafts provided sustained vertical moisture transport and contributed to the destabilization of pre-MCS$_S$ environment. On the other hand, the pre-MCS$_N$ environment was dominated by widespread downward motion, which inhibited environmental destabilization. The similar spatial characteristics for $\omega$ between WRF and RAP is encouraging and provides credence to our observational results.

Chapter 6 – Sensitivity of simulated systems to microphysical processes

6.1. Overview of microphysical sensitivity experiment designs

1. MCS strength and microphysical processes: An overview

The difference in the radar-observed microphysical properties indicated that MCS$_S$ featured more robust ice and mixed phase microphysics. Ice microphysics has long been identified to be critical in determining both the radar structure (e.g., Braun and Houze 1994) and internal circulation (e.g., Yang and Houze 1995; van den Heever and Cotton 2004; Adams-Selin et al. 2013a) of convective systems. Recent work by Adams-Selin et al. (2013b) identified a high sensitivity for various convective system characteristics (e.g., 10-m wind speed, system-
propagation speed, organizational pattern) to different microphysics schemes and further attributed these sensitivities to the parameterization for mixed phase particle properties.

Cold pool-environmental shear balance (Rotunno et al. 1988) provided an attractive framework on which the response of convective storm characteristics to changes in ice microphysical could be interpreted. Ice/Mixed phase process may impact the balance by adjusting the cold pool strength. Recent idealized studies by Adam-Selin et al. (2013a,b) showed slower cold-pool accumulation when high-density graupel. Cold pool accumulation was hindered in these experiments since the graupel descended too quickly, which limited the latent cooling rate. Substantial uncertainties exist on this particular relationship alone, however. For example, van Weverberg (2012) indicated stronger cold pools when rimed particle drop size distributions are slanted towards high-density particles while van den Heever and Cotton (2004) showed weaker cold pools instead. These results illustrate the need to revisit this issue under a variety of environments and model settings.

Rimed particle characteristics aloft can also impact features other than cold pools. Specifically, Yang and Houze (1995) showed that weaker rear inflow and midlevel low-pressure perturbation within simulated MCS could be achieved simply by implementing a separate hail category. The larger fall speed of hail was shown to be a main driver of these changes. Properties other than the strength of RTF flow also seem to be sensitive to rimed particle characteristics. For example, a postponement of RTF descent could be expected for systems with heavier graupel. These systems also featured less rearwardly-tilted updrafts and developed bow echoes at a slower rate (Adams-Selin et al. 2013).
The above studies suggest that the features most sensitive to ice/mixed phase processes are
cold pools and RIJs. The impact of cold pools on the durability and strength of convective
systems is generally well-documented (e.g., Dawson et al. 2010). The impact of RIJs is, in
contrast, comparatively less understood. Past studies on the impact of RIJs on MCSs (e.g.,
Xue et al. 2017) generally followed the dynamical argument put forth by Weisman (1992),
where the vorticity associated with RIJs partially neutralized the cold pool vorticity and
created more upright updrafts. The observed correlation between RIJ expansion and robust
mixed phase processes (Section 4.3) suggests that other changes caused by RIJs will also
need to be considered. Of particular relevance to this study is the so-called “hydrometeor
recirculation” (Siegel and van den Heever 2013), where mid-level updrafts are strengthened
as rimed particles re-entrained into the updrafts by the RIJs (Fig. 6.1.1).

2. Sensitivity experiment designs

(a) Sensitivity tests on hydrometeor size-sorting (Group I)

The first set of sensitivity tests (Group I) in this study concerns the ability for hydrometeors
to sediment at different rates based on their sizes (“size sorting”). The size sorting
phenomenon is represented in multi-moment microphysical schemes by the different
terminal velocities for hydrometeor number concentration ($V_N$) and mixing ratio ($V_q$).

$$V_N = \int_0^\infty \frac{\gamma c D^d n(D) dD}{\int_0^\infty n(D) dD} = \gamma c D^d \frac{a^u v_r f(v_{\mu+d})}{\alpha (v_{\mu+d})^n}$$ (Eq. 6.1.5)

$$V_q = \int_0^\infty \frac{\gamma c D^d m(D) n(D) dD}{\int_0^\infty m(D) n(D) dD} = \gamma c D^d \left[ \frac{\Gamma(b+\mu+d)}{\alpha^b+\mu+\frac{a^u v_r f(v_{\mu+d})}{\alpha (v_{\mu+d})^n}} \right]^{-1}$$ (Eq. 6.1.6)

The parameters $c$ and $d$ are the constants in the power law relationship for terminal velocity
($V(D) = \gamma c D^d$; $\gamma = (\rho_0/\rho_a)^{0.5}$). In a two-moment microphysical scheme, larger simulated

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particles will sediment at a quicker pace than smaller particles since $V_q$ is greater than $V_N$. However, two-moment schemes often overestimate size sorting without mitigating adjustment procedures (Mansell 2010).

The most important experiment in Group I is the GNSS (Graupel-no-size-sorting) experiment. In the GNSS, $N_g$ and $q_g$ are set to sediment under $V_q$, which caused the graupel fall speeds to be invariant to particle size. The other experiments in this group include RNSS (Rain-no-size-sorting), where rainwater (graupel) size sorting is inhibited (permitted) in the model; and GRNSS (Graupel-rain-no-size-sorting), where the size sorting for both graupel and rainwater are prohibited.

(b) Sensitivity tests on fall speed-diameter relationship (Group II)

The second group of sensitivity experiments in this group is related to the power law ($ycD^d$) for terminal velocity. The different parameters values used in the power-law equations between the experiments (Table 6.1.2) caused varied rimed particle terminal velocity change rate (Fig.6.1.2a). The parameters used in the FULL and HDGR experiments follows the Mansell et al. (2010) relations while Ferrier (1994) relations was used in the FERR. Rimed particle density in the FULL simulation was set to be $500 \text{ kg m}^{-3}$ while rimed particle density was set to be $\rho_{HDGR} = 900 \text{ kg m}^{-3}$ in the HDGR simulation. The terminal velocities for rimed ice particles for the HDGR are consistently larger than the FULL (Fig.6.1.2a). On the other hand, graupel terminal velocity increased at a slower rate for larger particles in the FERR (Fig.6.1.2a).
(c) Sensitivity test on rimed particle type (Group III)

Rimed particles in Group I and II simulations are all categorized as graupel. In the HAIL experiment, we examined this assumption by forcing all the rimed particles to be categorized as hail with \( \rho = 900 \text{ kg m}^{-3} \). While the particle densities are identical between the HDGR and the HAIL, different drag coefficients (\( C_D \); Table 6.1.2) between the assumed fall-speed power-law equations for hail and graupel increased the terminal velocity for the HAIL (Fig.6.1.2b).

6.2. Sensitivity experiments: Overview of simulated systems

1. Reflectivity structures

(a) FULL simulation (baseline experiment)

The FULL-simulated MCSs during the mature phase (Fig.6.2.1b) compares favorably to the observed MCSs (Fig.6.2.1a). Notably, the MCSs “transition zone” (Braun and Houze 1994) was successfully simulated by the FULL, indicating that the FULL ice/mixed phase hydrometeor properties were similar to observations.

(i) Convective region

The line-averaged reflectivity cross section (Fig.6.2.2b) along a NW-SE cross section (black line in Fig.6.2.1) indicated that large reflectivity values for the FULL were primarily concentrated near the convective edge, which compared well with observations (Fig.6.2.2a). While the simulated reflectivity values were slightly weaker than the observed values, this underestimation was likely caused by averaging rather than indicating weaker convection.
(ii) **Transition zone**

The FULL simulation was successful in creating a region with weaker reflectivity values between the convective and stratiform regions. Compared to the observations, however, the reflectivity minimum was weaker in the FULL, which likely indicated slight biases in the rimed particle fall-speed parameterization.

(iii) **Stratiform region**

The system stratiform region was not as successfully captured in the FULL since the simulated reflectivity values decreased rather than increased below the melting level. Potential reasons for this discrepancy include:

- Prevalence of slower-falling, smaller rain drops that are harder to grow by coalescence.
- Enhanced evaporation due to increased hydrometeor residence time.

**(b) The sensitivity of reflectivity pattern to hydrometeor size-sorting (GNSS, RNSS, GRNSS)**

During the system mature phase, the GNSS-simulated MCSs (Fig.6.2.3a,b) was structurally dissimilar to the FULL as it lacked transition zones. Higher reflectivity values in this system also located to the rear of the system rather than system edge.

Line-average cross section for the GNSS (Fig.6.2.3a) showed bright band signature below the melting level in the system rear while the reflectivity was weaker near system edge. The weak system edge reflectivity will be shown to be related to the lack of large graupel

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13 Subjectively determined to be the time when the simulated MCS reflectivity was most severe and most resembled the observed MCS bow echo.
sedimentation immediately below the updrafts.

The reflectivity structure for the RNSS (Fig.6.2.4a,b) was similar to the FULL, which pointed to the importance of rimed particle size sorting in creating the substantial inter-model differences in Group I. The above statement could be confirmed since the line-averaged reflectivity cross-section for the GRNSS (Fig.6.2.5b) was generally similar to the GNSS system.

(c) The sensitivity of reflectivity pattern to fall speed-diameter relationship (HDGR, FERR)

The horizontal structure of the HDGR system was similar to the FULL/RNSS/observations. Line-averaged cross section for the HDGR (Fig.6.2.7a), however, showed a less vertically-developed convective region and a wider trailing stratiform. For the FERR, the simulated system consisted of multiple convective cores within a quasi-linear system (Fig.6.2.6b). Compared to the other experiments, the FERR featured strong reflectivity values (>35 dBZ) that extended ~1-2 km higher while reflectivity was weaker below the melting level (Fig.6.2.7b). This probably meant that rimed particles were primarily transported upward by the updrafts instead of falling below the melting level.

(d) The sensitivity of reflectivity pattern to the choice of rimed particles (HAIL)

While both HDGR and HAIL featured greater terminal velocity increase rate with diameter, the HAIL system shared more similarities with the FERR system (Fig.6.2.8) since both systems exhibited narrow linear structures. Their structural similarities were likely accidental: the system narrowness could be attributed to increased rimed particle upward transport and weaker horizontal FTR rimed particle transport for the FERR and FERR, respectively.
2. Microphysical characteristics: Hydrometeor distributions

(a) FULL simulation and Group I sensitivity tests

Similar to the reflectivity structures discussed in Section 6.2.1, the GNSS and GRNSS hydrometeor distribution (Fig.6.2.9) differed significantly from the FULL. Due to smaller \( q_g \), the GNSS and GRNSS systems were dominated by slower-falling, lower-density snow particles. For the FULL and RNSS, greater riming rate increased the \( q_g \) substantially, as shown by the considerable supercooled rain and cloud water near system edge (Fig.6.2.9a,c).

The lack of size sorting for the GNSS and GRNSS also impacted the hydrometeor distribution aloft. Heavier rimed particles could quickly fall out of the updrafts if size sorting was allowed (i.e. FULL/RNSS), which would generate preferential graupel sedimentation near the system edge. This process could not occur in the GNSS/GRNSS since the graupel terminal velocities in these experiments were invariant to particle size. As a result, the distance between the system edge and graupel sediment location increased relative to the FULL/RNSS.

(b) FULL simulation and Group II sensitivity tests

Compared to the FULL, the FERR featured more vertically distributed graupel while \( q_r \) decreased by \(-0.4 \text{ g kg}^{-1}\) (Fig.6.2.10b). Decreased \( q_r \) was indicative of decreased graupel fallout as most graupel were transported upward. Compared to the GNSS, the riming process for the FERR was more robust since the \( q_g \) for the FERR aloft.

For the HDGR, wide stratiform region and weak reflectivity values within transition zone...
illustrated the dominant role of low-density particles. Increased mobility of ice-phase hydrometeors was confirmed in Figure 6.2.10a since more snow particles were produced above the melting level. These particles could travel a greater distance before sedimentation, thereby creating a wider stratiform region. These was also a distinct lack of snow or graupel mass below the melting level within the transition zone, which created the simulated low reflectivity values.

Despite the lower $q_s$ in the HDGR convective region, the $q_r$ below the melting level was only slightly smaller than FULL. This result is counter-intuitive considering the Group I result. The significant rearward tilt of the HDGR updrafts could potentially explain this result: Extreme updraft slope in the HDGR limited the vertical extent of supercooled cloud water above the melting level (the maximum height of 0.3 g kg$^{-1}$ contour for the cloud mixing-ratio in the HDGR lowered by $\sim$1 km compared to the FULL), which effectively inhibited riming growth while elevated the importance of auto-conversion and collection processes in rainwater production. The net effect of these microphysical changes reduced $q_s$ considerably while maintaining $q_r$ to values comparable to the FULL and RNSS.

(c) FULL simulation and Group III sensitivity tests

Consistent with the discussion in Section 6.2.1, $q_h$ in the HAIL line-averaged mixing ratio cross section (Fig.6.2.11) was the most horizontally concentrated of the seven experiments discussed in this section.

While the simulated systems in the FERR and HAIL are similarly narrow, $q_r$ was larger for the HAIL (Fig.6.2.11). The rainwater mass enhancement was caused by greater rimed
particles fallout below the melting level that provided excess rainwater.

3. Accumulated Precipitation

The inter-model differences for the accumulated precipitation from 0300 to 0700 UTC (Fig.6.2.12), which varies from ~33 mm (HDGR) to ~62 mm (the RNSS), were quite large. Preferential larger graupel sedimentation for the FULL and RNSS lowered the hydrometeor residence time, thus increasing the accumulated precipitation. For the other simulations, accumulated precipitation was lowered either by enhanced graupel lifting (FERR) or due to the dominance of slow-falling snow particles aloft (GNSS/GRNSS), which was detrimental to precipitation accumulation (Fig.6.2.9b,d) due to increased residence time.

The accumulated precipitation for the HDGR and HAIL were peculiar and thus warranted further discussions. The $q_r$ in the convective region was large for both (Fig.6.2.10a, Fig.6.2.11), yet the rainwater was unable to effectively convert to precipitation (Fig.6.2.12e,g). Increased (decreased) tendency for rainwater formation via warm rain processes (melting) in the sloped HDGR updrafts reduced the average diameter for rain drops below the melting level (Fig.6.2.13). As such, rainwater evaporation would be stronger below the melting level, limiting the system’s precipitation efficiency. Similarly, the smaller hail mass (Fig.6.2.14a-c) in the also limited the accumulated precipitation despite the high intrinsic terminal velocities for hail.

4. Near-surface temperature

Simulated temperature at the lowest model level was compared in Figure 6.2.15. The cooling
was stronger for FULL and RNSS due to increased rimed particle melting. Stronger water loading for these systems also favored intense near-surface wind, which increased the rain drop breakup-evaporation potential\textsuperscript{14} within the low RH layer near ground (Fig.6.2.16b). This process indirectly intensified the near-surface cooling and was vital for cold pool accumulation since only a negligible amount of graupel could reach the lowest 1 km (Fig.5.2.14b,c). On the other hand, the surface cooling was weaker for the GNSS and GRNSS due to weaker evaporation, as indicated by the smaller rainwater concentration decrease (Fig.6.2.16c) and weaker latent cooling (Fig.6.2.16a).

The microphysical cooling processes were weaker in the FERR when most graupel were transported upward, thus creating a cold pool warm bias compared to the other simulations. On the other hand, the cold pool was stronger for the HDGR due to the smaller mean rain drop diameter in the convective region (Fig.6.2.13).

### 6.3. Impact of size sorting on simulated systems – Results from Group I

#### 1. Evolution of updraft statistics

Updraft statistics for three separate phases of the systems’ life cycle: developing phase, mature phase, and weakening phase are compared in Figure 6.3.1. Updraft intensities during the mature and weakening phases were generally weaker due to enhanced downward

\textsuperscript{14} Notice the substantial rain concentration decrease (Fig.6.2.16c) and increasing mean-mass rain diameter (Fig.6.2.13) in the lowest 2 km layer, both indicating a loss of smaller rain droplets by evaporation.
buoyancy pressure gradient force (DBPGF; Parker 2010) as updraft tilt increased. However, the updraft weakened at a substantially slower rate for the FULL, which indicated graupel size sorting did have an impact in system strength evolution. During the weakening phase, the GNSS updrafts slightly strengthened compared to the FULL updrafts as new convection developed southwest of the system. Contrastingly, the updraft statistics for the RNSS was similar to the FULL during mature and weakening phases. This also confirmed our conclusion from Section 6.2 that on the importance of mid-level microphysics. This was probably not unexpected, considering the large inter-model latent cooling discrepancies near the melting level (Fig.6.2.16a).

The importance of graupel size sorting in limiting updraft weakening was further supported by the GRNSS (Fig.6.3.1c), where the updraft evolution shared more similarities to the GNSS than the RNSS. Due to the similarity between GNSS and GRNSS experiments, only the results from the FULL, GNSS, and RNSS will be examined in the remainder of this paper.

2. Evolution of updraft vertical structure

In this section, the CFAD (Contoured-frequency by Altitude Diagrams) technique was utilized to expand upon the results of the previous section. Specifically, we would like to identify the vertical layer where these updraft differences occurred.

The FULL CFADs for stronger updrafts (\(w > 2 \text{ ms}^{-1}\)) during the system developing phase

\footnote{Due to considerable time lag between FULL and RNSS mature phase, FULL updraft statistics simulation during time period \((t)\) are compared to the corresponding GNSS statistics during a different period, defined by \((t+1\text{hr})\). (e.g., 03:30-04:30 UTC [FULL] is compared to 04:30-05:30 UTC [RNSS])}
Fig. 6.3.2a shows rapid expansion of updraft distribution above surface, which represented the latent heat release as vapor condensed. The strongest updrafts, however, could be found above the melting level, where robust riming and vapor deposition could be expected to intensify the updrafts.

Due to the overall similarities between the CFADs for different experiments (especially during the developing phase), full updraft CFADs will not be shown and individual frequency contours will be compared instead for the sensitivity experiments. The frequency contour utilized in the analyses therein was chosen to be 0.135 since it encompassed the updraft strength bracket of 4-5.5 m s\(^{-1}\), where the differences between experiments was the largest during the developing-mature phase transition (Fig. 6.3.1).

For the GNSS, updrafts were generally weaker than the FULL during the developing phase, especially above the melting level. These differences exacerbated during the mature phase (Fig. 6.3.2b), where the FULL updrafts were stronger than the GNSS by up to ~2 m s\(^{-1}\) above 5 km AGL and near the melting level. The updraft characteristics in low troposphere were generally similar. The mid-level updraft strengthening was supportive of our hypothesis that “hydrometeor recirculation” process played a role in the IOP30 system maintenance since the availability of smaller rimed particles to be re-entrained into the updrafts will potentially be lower without size sorting.

For the RNSS system, the 0.135 updraft frequency contour was similar to the FULL counterpart apart from a narrow layer near 7 km AGL during both the developing (Fig. 6.3.3a) and mature phase (Fig. 6.3.3b). This result is encouraging since it suggested that the mid-level
process that maintained mid-level updrafts were also at work in the RNSS system. The RNSS updraft, on the other hand, was consistently weaker than the FULL below the melting level. We can, therefore, conclude that the updraft strength variations for Group I system manifested more strongly in the mid-levels than the lower levels.

3. Potential link between MCS strength and rear-inflow jets (RIJs)

A comparison on the kinematic structures of the simulated systems showed that experiments with stronger mid-level updrafts (Fig. 6.3.4a,c) tended to have stronger 4-km$^{16}$ rear-to-front flow. The rear-to-front flow in these experiments could also extend further into the convective edge.

Figure 6.3.5 shows the average vertical cross section taken along a 30km sector with the strongest simulated reflectivity and rear-to-front wind speeds. The FULL system (Fig. 6.3.5a) was characterized by a descending RIJ with a $\sim 17$ m s$^{-1}$ wind maxima near the convective edge (Fig. 6.3.5a). The RIJ remained elevated and did not descend before it reached the convective region, which was undoubtedly favorable for hydrometeor recirculation process. Contrastingly, the GNSS RIJ was weaker, horizontally limited, and descended far to the system’s rear compared to the FULL.

The RNSS circulation (Fig. 6.3.5c) was again similar to the FULL. Its FTR flow was slightly more tilted while its RIJ descended earlier than the FULL. The strength and horizontal extant

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$^{16}$ This altitude was chosen as a representative layer at which the strongest mid-level updraft strength differences were found.
of the RNSS RIJ were much closer to the FULL than GNSS, however.

4. Microphysical-dynamical contribution of stronger rear inflows in FULL/RNSS

In this section, we will attempt to understand why certain experiments (i.e. FULL, RNSS) feature stronger and more expanded RIJs than the others. Some crucial differences, including (a) slightly larger tilt for FULL/RNSS updrafts, and (b) lack of divergent flow pattern for the GNSS, emerged during the developing phase that could potentially cause the initial RIJ strengthening for the FULL/RNSS.

Latent cooling near the melting level was weaker for the GNSS (Fig.6.3.7b) since its smaller updraft tilt caused particles to sediment in locations that were unfavorable for different cooling processes. This could be shown by examining the humidity conditions within the main graupel sedimentation region for different experiments (Fig.6.3.9). Relative humidity with respect to ice and water are defined as follows,

\[
\begin{align*}
RH_{\text{water}} &= \frac{\sigma}{\sigma_s(T)} \times 100 \\
RH_{\text{ice}} &= \frac{\sigma}{\sigma_i(T)} \times 100
\end{align*}
\]  
(Eq. 6.3.1)

where \( \sigma \) is the vapor partial pressure, \( \sigma_s(T) \) and \( \sigma_i(T) \) are the saturated vapor pressure over water and ice.

As shown in Figure 6.2.16e, graupel in the FULL/RNSS fell as far as 2 km AGL (\( \sim 9.7^\circ \text{C} \)) before they fully melted. The RH_{ice} profiles for these two experiments (Fig.6.3.9b) showed

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\(^{17}\text{Defined as the region with } q_g > 0 \text{ g kg}^{-1} \text{ and with latent cooling rate greater than } 0.5 \text{ K (5 min)}^{-1} \text{ below the melting level.}\)
sub-saturated air with regard to the un-melted ice between ~-4°C and ~10°C. On the other hand, the decrease rate of RH\textsubscript{water} with increased temperature was considerably smaller than RH\textsubscript{ice}: RH\textsubscript{water} decreased by 5.76% from -4°C to 5°C, while the corresponding RH\textsubscript{ice} decrease rate was 15%. These results indicated that stronger latent cooling above 2 km AGL in the FULL/RNSS systems were primarily induced by graupel melting or sublimation, while rainwater evaporation was more important below 2 km. For the GNSS system, RH\textsubscript{ice} in the graupel fallout region was ~10% higher (Fig.6.3.9b) than for the FULL/RNSS while RH\textsubscript{water} remained close to saturation from 0-10°C (Fig.6.3.9a). The higher RH was consistent with the weaker latent cooling for the GNSS since it inhibited both evaporation or sublimation processes. To conclude, latent cooling in the FULL/RNSS systems were bolstered by enhanced graupel melting and sublimation below the updraft (Fig.6.3.7). Conversely, latent cooling in the GNSS system was depressed due to greater humidity (Fig.6.3.8-9) below the updraft.

We will argue here that the dynamical response to the differences in latent cooling strength documented above was the primary cause for the initial RIJ intensification in the FULL/RNSS. This response could be understood under a pressure perturbation framework.

\[
\frac{1}{\rho_0} \nabla^2 p' = -\mathbf{V} \cdot (\mathbf{V} \cdot \nabla \mathbf{V}) + \frac{\partial B}{\partial z} \quad (Eq. 6.3.4)
\]

\[
\begin{align*}
\frac{1}{\rho_0} \nabla^2 p'_B &= -\mathbf{V} \cdot (\mathbf{V} \cdot \nabla \mathbf{V}) \\
\frac{1}{\rho_0} \nabla^2 p'_B &= \frac{\partial B}{\partial z}
\end{align*}
\]

The first (second) term of the right-hand side of Eq.6.3.4 represents the dynamical (buoyancy) contribution to total system pressure perturbation. The larger buoyancy gradient (Fig.6.3.10b) in the FULL system could, theoretically (i.e. Eq.6.3.4), induced stronger low-pressure
anomaly that could expand and strength the RIJs.

5. Thermodynamical and kinematic structures during mature phase

While the above findings were encouraging, they are primarily based on the system’s developing phase and might not be relevant during mature phase. We therefore repeated the above analyses for the mature phase to ease this doubt.

The buoyancy structures during mature phase (Fig.6.3.11) showed that experiments with graupel size sorting (Fig.6.3.11a,c) tended to feature stronger positive (negative) buoyancy above (below) the melting level. The negative buoyancy in these experiments was also located closer to the system edge. The enhanced buoyancy largely correlated well with the stronger latent heating and cooling rates for these systems (Fig.6.3.12a,c). These changes increased the buoyancy gradient near the system edge considerably.

The horizontal distance between updrafts and latent cooling below the melting level was larger for both the GNSS (Fig.6.3.12b) and GRNSS (Fig.6.3.12d). This misalignment relaxed the buoyancy gradient and created weaker and less horizontally expanded RIJs through weaker low-pressure anomalies. Weaker RIJs could adversely impact the efficiency of hydrometeor recirculation and weaken the mid-level updrafts (Section 6.3.1-2).

6. Microphysical characteristics during MCSs mature phase

The microphysical structures for the simulated systems could be studied with mean-volume diameter ($D_{v,x}$),

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\[ D_{v_x} = \left( \frac{6\rho_a q_x}{\pi \rho_x N_{T_x}} \right)^\frac{1}{3} \]  

(Eq. 6.3.5)

where \( \rho_a \) (\( \rho_x \)) represents air density (density of specific hydrometeor type), \( q_x \) is the hydrometeor mixing ratio, and \( N_{T_x} \) is the total number concentration. \( D_{v_g} \) for the FULL was larger near system edge and increased towards ground, which was consistent with graupel size sorting (Fig.6.3.13a). The spatially uniform \( D_{v_g} \) for the GNSS, on the other hand, matched the prohibition of graupel size sorting inherent in the GNSS (Fig.6.3.13b).

The graupel diameters PDFs above ~2 km AGL\(^{18} \) (Fig.6.3.14) resembled gamma distribution for the FULL/RNSS while the PDFs were slanted towards larger particles for the GNSS/GRNSS. A potential explanation for this bimodality could be provided following Jensen et al. (2018): Under similar environmental moisture conditions, rimed particle formed in weaker updrafts will be larger on average but less numerous. Smaller overall \( D_{v_g} \) in stronger systems could be attributed to secondary ice production near the melting level, which could rapidly generate a large amount of small rimed particles. A comparison of \( N_{\text{graupel}} \) statistics (Fig.6.3.15) and average \( N_{\text{graupel}} \)-ambient temperature joint distribution (Fig.6.3.16) reveal two critical features that supported our hypothesis:

- Higher \( N_{\text{graupel}} \) in the FULL/RNSS systems (Fig.6.3.15).
- Extra graupels were most concentrated between -3 °C and -10 °C (Fig.6.3.16a), which compared well with the main temperature range for rapid riming growth of rime splinters, -3 °C ~ -8 °C (Hallet and Mossop 1974).

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\(^{18} \) This height was chosen based on the predominate graupel melting at this height.
These microphysical inferences could be confirmed by examining the vapor increasing/decreasing tendency near the updrafts. For the FULL, graupel melting likely dominated the overall cooling budget since there was little vapor enhancement in areas with latent cooling below the updrafts (Fig.6.3.17a). Vapor increase was concentrated in a narrow vertical band ~18 km behind the updrafts, which featured greater graupel sublimation. Elsewhere, vapor increase in the lowest 1.5 km was caused by rainwater evaporation in sub-saturated region (not shown). For the GNSS, vapor enhancement was weaker yet distributed over a broader vertical region (Fig.6.3.17b). The weaker overall cooling reflected weaker melting when fewer graupel formed aloft. The GNSS also featured weaker evaporation\(^1\) due to the ineffective *indirect* process noted in Section 6.2.3.

### 7. Brief summary on the results from Group I

Through the above analyses, it was found that graupel size sorting was beneficial to MCS maintenance through a positive feedback loop, summarized below.

1. Size sorting enabled a portion of heavier graupel to *fall in locations closer to updrafts*.
2. Latent cooling would shift towards the updrafts due to graupel melting (Fig.6.3.14).
3. Negative $P'_B$ was generated below updrafts due to heightened buoyancy gradient (Fig.6.3.10), which favored RIJ strengthening and expansion.
4. Mid-level updrafts strengthened (Fig.6.3.2) due to “hydrometeor recirculation”.
5. Latent heating and cooling were stronger under stronger updrafts owing to the rapid riming growth of secondary ice (Fig.6.3.13-18) and graupel melting, representatively.

Latent cooling was also stronger near the ground as a result of raindrop evaporation in

\(^1\) Notice the smaller vapor increase in the lowest 1.5 km compared to FULL.
6. Both processes in (5) exerted positive impacts on the continuation of this feedback loop.

6.4. Impact of rimed particle sedimentation characteristics on simulated systems – Results from Group II & III

1. Sensitivity to terminal velocity-diameter relationship

In Section 6.2, we found out that, for the Group II and III simulations:

- More rimed particles were transported upward (Fig.6.2.10b) in the FERR.
- Greater rainwater evaporation in the HDGR (Fig.6.2.13, Fig.6.2.15e) generated stronger cold pools (Fig.6.2.15e).
- Decreased graupel mass within the upshear-tilted updrafts in the HDGR due to insufficient supercooled drops aloft (Fig.6.2.10a).

The effects of these microphysical changes will be examined further in this section.

(a) Statistical analysis on updraft strength

During the developing phase (Fig.6.4.1a), the HDGR updrafts were weaker than the FULL both below and above the melting level. The weaker updrafts aloft were associated with the stronger updraft tilt (Fig.6.4.2) while the weaker updrafts in the lowest 1-2 km implied weaker cold pools (Xue et al. 2017). The HDGR system’s mid-level updrafts was even weaker compared to the FULL during the mature phase (Fig.6.4.1b), which was consistent with our inference in Section 6.2.2(b) on the lack of rime growth in the HDGR. There was, however, a slight strengthening tendency for the updrafts in the lowest 2 km. The system cold
pool was stronger during this period as a consequence of increased evaporation below the melting level (Fig.6.2.13). Finally, the 0.135 frequency contour was similar for the HDGR and the FULL during weakening phase. Examination of the original CFADs (not shown) shows that the lack of riming had depleted the strongest updrafts and created a narrower updraft distribution compared to the FULL.

During the initial developing phase, the 0.135 frequency contour for the FERR and the FULL (Fig.6.4.3a) stayed relatively close each other below ~6 km AGL. This pattern remained unchanged during the mature phase (Fig.6.4.3b). The similar CFADs between the GNSS and the FERR aloft indicated that the FERR would likely weaken at a quicker rate than the FULL. Indeed, there was a notable decrease in the potential for the strongest updrafts (i.e. 0.01 frequency contour) during the weakening phase for the FERR (Fig.6.4.3d). The role of mid-level microphysical processes (e.g., graupel recirculation) on the weakening of the FERR system will be explored in later sections. Below the melting level, however, the 0.135 contour shifted towards stronger updrafts compared to mature phase, which demonstrated a slower cold pool accumulation rate than the FULL when the Ferrier fall speed-diameter relationship was used.

(b) Line-averaged two-dimensional storm circulation analysis
The line-averaged circulation for the HDGR (Fig.6.4.4a) featured highly tilted FTR flow and weak, quickly-descending RTF inflow. These results compared well with Adams-Selin et al. (2013), which associated the quicker-descending RIJs with the systems with larger stratiform regions. The quick descent and the surface-based nature of the system RIJ (x~135 km in Fig.6.4.4a) could negatively impact the maintenance of the HDGR system by increasing the
updraft tilt (Weisman 1992; Fig.6.1.1). Following the discussions on Group I, the less elevated RTF inflow (Fig.6.4.4a) also hindered system maintenance through hydrometeor recirculation. This could be confirmed separately by the high correlation between $q_g$ and $N_{Tg}$ distributions near the ascending inflow layer ($x=120-140$ km in Fig.6.4.5).

Compared to the FULL, the FTR flow in FERR system was minimally tilted while the RTF flow was also weaker (Fig.6.4.4b). RIJ weakening in this experiment was induced by a rearward shift in latent cooling location that somewhat impeded the hydrometeor recycling process.

(c) MCS thermodynamical structure and rimed particle characteristics during mature phases

The latent heating maxima for the HDGR were located below the melting level (Fig.6.4.6a), which indicated lesser importance for mixed phase processes. Limited graupel mass aloft and suppressed graupel fallout (small $D_{vg}$ in the updraft) also weakened latent cooling near system edge (Fig.6.4.6a). On the other hand, latent cooling was stronger in the system rear due to the melting of low-density particles (Fig.6.2.10a).

The overall latent heating and cooling distributions for the FERR (Fig.6.2.10b) generally resembled the FULL. There were, however, some differences that could weaken the FERR system, including:

- Weaker latent cooling below ~3.8 km
- Greater horizontal separation between the latent cooling and heating region near the
melting level (Fig. 6.2.10b), which was consistent with weaker graupel size sorting in the FERR\textsuperscript{20}.

These changes weakened the buoyancy gradient (Fig. 6.4.7b) and resulted in weaker and less expanded RTF flow.

\textit{(d) Summary for Group II systems}

Based on above analyses, a correlation between quicker system weakening and \textit{weaker} rimed particle size sorting could be established. This correlation seems to arise from the contraction and weakening of the system RIJs associated with suppressed heavy particle fallout below the updrafts. The results were less conclusive for experiment with \textit{stronger} size sorting (HDGR), however. For this experiment, updraft tilt replenished the convective region with smaller rimed particles and therefore negated the positive impact of stronger size sorting.

\textbf{2. Sensitivity to dominant rimed particle type}

\textit{(a) Statistical analysis on updraft strength}

A comparison between the 0.135 updraft frequency contours for the FULL and HAIL (Fig. 6.4.8) reveals several differences, which are summarized as follows:

- Substantially weaker low-level updrafts for the HAIL during the developing phase (Fig. 6.4.8a), which rapidly intensified afterwards (Fig. 6.4.8b-c).
- The mature phase updrafts were weaker aloft than the FULL by an average of \textasciitilde 1 ms\textsuperscript{-1} while updraft strength was close to the FULL near the melting level.

These differences suggest that the HAIL system was dominated by low tropospheric

\textsuperscript{20} Notice the similar $q_e$ and $N_f/e$ distributions below the updrafts (Fig. 6.4.5b).
processes and its cold pool accumulation rate was higher than the FULL. The similar FULL and HAIL CFADs above the melting level during the mature and weakening phases further revealed that mid-tropospheric processes (i.e. hydrometeor recirculation) in the HAIL system was probably only slightly weaker than the FULL, despite the smaller rimed particle mixing ratio aloft (Fig.6.2.11).

(b) Cross-section analysis on storm thermodynamical, microphysical, and circulation structures

Compared to the other six experiments in this study, the defining characteristics of the HAIL system is the rapidity of cold pool accumulation. Under low-shear conditions, the strong cold pool in the HAIL system would be detrimental to system intensification under the RKW theory. The significant rearward tilt of the updrafts (vectors in Fig.6.4.9a) also meant that hail particles near the system edge would be smaller than system rear (Fig.6.4.10). However, due to the narrowness of the HAIL system (partially contributed by the higher fall speed), the RTF inflow could still transport a considerable amount of graupel back towards the updrafts (Fig.6.4.10).

As a result, the latent heat release (Fig.6.4.9a) was weak near the edge of the HAIL system but was stronger near the region where the RTF flow collided with the convective updrafts. The latent heat release adjusted the buoyancy distribution near the convective region (Fig.6.4.11a) in the way that generated negative $P_B'$ that was stronger than the FERR near the melting level (Fig.6.4.11b).

The above discussion indicated that the circulation for the HAIL system allowed greater
system maintainability due to considerable RTF inflows into convective updrafts. The $q_v$ and $q_c$ tendencies near the HAIL updrafts (Fig.6.4.12) show that the latent cooling maximum in the lowest 1 km (1.5-2.5 km) roughly correlated with positive vapor tendency (Fig.6.4.12a), indicating the crucialness of evaporative cooling in cold-pool accumulation. Closer to the melting level, we note a high spatial correspondence between latent cooling and positive vapor tendency, which illustrated elevated contribution of sublimation to the overall latent cooling budget. Finally, hydrometeor recirculation in this experiment could be verified as there was a greater propensity for $q_v$ and $q_c$ sinks to occur in locations where the RTF flow encountered the updraft (Fig.6.4.12b,c).

6.5. Dynamical inferences on simulated MCS$_S$

Brief dynamical discussions on the simulated MCS$_S$ pressure perturbation and vorticity fields will be discussed to identify potential contributors to the expansion of system RIJs and its impact on system maintenance. For brevity, we will only discuss the results for the FULL and GNSS systems.

1. Diagnostical analysis on pressure perturbation

(a) Analysis method

The two Poisson equations for dynamical pressure perturbation ($P_D^*$) and buoyancy pressure perturbation ($P_B^*$) were solved numerically by applying Neumann boundary conditions.

(b) Analysis results and discussion
Diagnosed $P_B'$ structures relative to the convective updraft are shown in Figure 6.5.1. The amplitude of the $P_B'$ were similar to Lebo (2018), which indicates that the results calculated with our simplified method were reasonable. For the FULL (Fig.6.5.1a), a $P_B'$ dipole formed behind the convective updraft in response to the buoyancy structure near system edge. The negative $P_B'$ expanded horizontally to the rear of the system due to snow melting.

The impact of $P_B'$ distribution on the system circulation could be analyzed by computing the pressure gradient force associated with $P_B'$ (PGF$_P_B$), defined as follows,

$$PGF_{P_B} = -\frac{1}{\rho_0} \nabla P_B'$$ (Eq. 6.5.1)

The line-averaged horizontal component of the PGF$_P_B$ for the FULL (Fig.6.5.2a) behaved in wavelike manner in the forward anvil, possibly due to smaller scale features. The PGF$_P_B$ distributed more coherently near the convective region. The dynamical forcing associated with PGF$_P_B$ intensified the mid-level RTF flow near and behind the convective updrafts, which was beneficial in the strengthening and expanding the RIJ, and increasing the effectiveness of “hydrometeor recirculation”.

For the GNSS, the negative $P_B'$ (Fig.6.5.1b) shifted rearward compared to the FULL. Therefore, the RTF flow was less expanded and would not be able to re-transport a substantial amount of graupel back to the updrafts (Fig.6.5.2b).

Following Peters (2016), the dynamic pressure perturbation ($P_D'$) could be expanded to,

$^{21}$ The vertical component of the PGF$_P_B$ was not discussed in this study. The reason for this omission is because our primary research focus is on the generation and expansion of the RIJ. We do acknowledge a potential correlation between updraft amplitude and the vertical PGF$_P_B$, as discussed in Parker (2010) and Lebo (2018).
\[ \nabla^2 p_D' = -\rho_o \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial y} \right)^2 + \left( \frac{\partial w}{\partial z} \right)^2 - w \frac{d^2 \ln \rho_o}{dz^2} \right] - 2 \left( \frac{\partial u}{\partial y} \frac{\partial v}{\partial x} + \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} + \frac{\partial w}{\partial z} \frac{\partial v}{\partial x} \right) \] (Eq. 6.5.2)

For the FULL, the negative \( p_D' \) near the system updrafts (Fig.6.5.3a) could be attributed to the strong shear between the FTR and RTF flows (Szeto and Cho 1994). However, the \( p_D' \) amplitudes were only at \( \sim 1/10 \) of the corresponding \( p_B' \) near the convective updrafts, indicating that the impacts of \( p_D' \) on the system were likely small. The \( p_D' \) for the GNSS (Fig.6.5.3b) was even weaker than in the FULL and was therefore unimportant.

2. Diagnostical analysis on vorticity evolution

(a) Description on vorticity distribution

The vertical vorticity at 2.5 km AGL for the FULL (Fig.6.5.4a) shows cyclonic (C1) and anticyclonic bands (A1) paralleled to the system. However, C1 extended northward and formed counterrotating vortices (C2) perpendicular to the convective line with the A1, C1 and A1 formed a distinct vorticity couplet behind the developing bowing segment, which could dynamically enhance the RIJ (Fig.6.5.5a). The GNSS (Fig.6.5.4b) featured weaker counterrotating vortices (i.e. the anticyclonic circulation), which provided less dynamical forcing for the RIJ expansion (Fig.6.5.5b).

(b) Physical contributions to FULL vorticity structure

The vertical vorticity structure for FULL system (Fig.6.5.6a) exhibited high correlation between main convective updraft (downdraft) and the positive (negative) vorticity bands, which indicated that vorticity tilting and stretching were likely critical. The contributions of tilting and stretching terms to changes in vertical vorticity are given by,
\[
\frac{d\zeta}{dt} = -\left(\frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z}\right) - \zeta \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \quad (Eq. 6.5.3).
\]

The simulated \(\zeta\)-bands (i.e. A1, C1) were coincided with positive (negative) vorticity tilting and stretching terms (Fig.6.5.6c). These band-like features (Fig.6.5.6a) were related to environmental horizontal vorticity tilting by alternating internal vertical motions since the vorticity vectors were directed towards the system near these bands (Fig.6.5.6c). The horizontal vorticity vectors in the vicinity of C2, on the other hand, were directed to the southwest, which suggested that C2 was likely generated by the tilting of baroclinically-generated horizontal vorticity associated with MCSs cold pool.

The vertical vorticity structures along the cross-section AA’ illustrated positive \(\zeta\) near the ascending inflow layer and near the stratiform mesoscale updraft (Fig.6.5.7a). The positive \(\zeta\) also overlapped areas with positive vorticity tilting and stretching (Fig.6.5.6b), which attested to the importance of environmental vorticity tilting/stretching for our case. On the other hand, negative \(\zeta\) near surface likely formed in response to the cold pool negative buoyancy. Finally, the elevated negative \(\zeta\) near the melting level was formed by the tilting and stretching (Fig.6.5.6b) of the horizontal vorticity (Fig.6.5.7) above the RIJ core.

The horizontal vorticity distribution along AA’ (Fig.6.5.7c) revealed elevated negative (positive) vorticity maxima above (below) the RIJ core. The negative horizontal vorticity maxima roughly balanced the environmental vorticities aloft. In the low-levels, however, the negative vorticity associated with cold pools dominated the vorticity balance, thus enhancing the updraft tilt as a result.

(c) Physical contributions to GNSS vorticity structure
Following the above discussions, weaker counter-rotating vortices for the GNSS could potentially be attributed to the weaker RIJs and convective cold pools. The analyses on the GNSS vertical velocity and vorticity tilting/stretching structures (Fig.6.5.6b,d) indicated two features that were in agreement with the above hypothesis,

- Little changes in horizontal vorticity directions between environment and system rear.
- Weaker positive $\beta$ near the edge of the GNSS system.

The vertical velocity distribution at 2.5 km AGL for the GNSS system showed weaker updrafts near the system edge (Fig.6.5.6b), which inhibited the generation of positive $\zeta$ within the inflow layer through insufficient vorticity tilting/stretching (Fig.6.5.6d). Figure 6.5.6b also indicated weaker low-level downdrafts in the GNSS system rear, this could be attributed to the weaker latent cooling and water loading (Section 6.3). The negative vorticity maxima at the system rear weakened by $\sim 46\%$ compared to the FULL.

The vertical vorticity structure along the cross-section BB’ (Fig.6.5.8a) showed positive $\zeta$ within the convective updrafts due to positive vorticity tilting and stretching (Fig.6.5.8b). In contrast to the FULL, vorticity vector direction suggests that the negative $\zeta$ in the system rear was generated from environmental vorticity tilting rather than through baroclinic process (Fig.6.5.8a,b). This weakened the negative $\zeta$ compared to the FULL. Another feature that differentiated the two vorticity structures was a lack of elevated negative $\zeta$ in the rear of convective updrafts (Fig.6.5.8a). A comparison between the horizontal vorticity and system circulation structures along BB’ (Fig.6.5.8c) and AA’ (Fig.6.5.7c) indicate that this

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22 The amplitude of the negative $\zeta$ maximum was $\sim 3.35 \times 10^{-3}$ s$^{-1}$ for the FULL system. The corresponding negative $\zeta$ maximum for the GNSS system was $\sim 1.72 \times 10^{-3}$ s$^{-1}$. 

doi:10.6342/NTU201900175
difference was caused by a substantial weakening of the horizontal vorticities induced by the RTF inflow jet (RIJ). These two features inhibited the formation of the anticyclonic circulation, which could then impede the strengthening and expansion of RIJ. Furthermore, the horizontal vorticity structure along BB’ was dominated by negative vorticity in the convective system, which weakened the inflow layer ascent.

**Chapter 7 - Conclusion and future research routes**

**7.1. Summary of observational and simulation results**

In this study, we utilized the IOP30 case to address (a) the origin of cold pool variabilities for nocturnal convective systems, and (b) the contrast between sustaining tendency for CIs near system edge and the weakening tendency for CIs in the system rear. Observational analyses in Chapter 4 suggest that the system with more system edge CI (MCS$_S$) initially developed in an area with greater low-tropospheric moisture. Hourly RAP analysis revealed the role of frontogenetic transverse circulation in locally moistening (drying) the pre-MCS$_S$ (pre-MCS$_N$) environment. The resultant $\Delta Z_{LFC}$ difference contributed to the rearward CI location shift and the weakening of MCS$_N$. However, the environmental heterogeneity, while significant, was probably insufficient in explaining the observed system variability since (a) pre-MCS$_N$ CAPE increased above boundary layer by 0300 UTC (Section 4.1), and (b) MCS$_S$ would encounter areas with lower precipitable water along its path (Section 4.2). With the available radar observations, differences between the systems’ internal structures were evident. Specifically, the strength and extent of the RTF flow for MCS$_S$ was stronger both near the melting level and the ground than MCS$_N$. These kinematic characteristics could be beneficial to system maintenance through the hydrometeor recirculation process (Siegel and
van den Heever 2013). The dual-polarized parameter profiles (Section 4.3) showed signs of robust mixed phase process and increased convective characteristics for MCSs, which could potentially affirm the importance of ice microphysics in the longevity of system edge CI and in creating the observed system variabilities.

A microphysical modelling approach was adopted to evaluate the relative importance of two hypotheses proposed in Section 2.3. Since the pre-MCS instabilities were similar between the experiments, the impact of ice microphysics will be crucial if the inter-experiment differences were large. Graupel sedimentation was chosen to be the targeted ice microphysical process in this study due to its impact on the of the hydrometeor recirculation process (Section 6.1). Encouragingly, simulation comparisons in Chapter 6 revealed high sensitivities between updraft longevity and system circulation to graupel sedimentation. Specifically, the system RIJ was stronger and more expanded when rimed particles were allowed to sediment with different terminal velocities in accordance to their sizes. Changes in latent heating/cooling structures were determined to be the main culprit for this sensitivity. For these systems, preferential fallout and melting of larger rimed particles below the updrafts strengthened the negative buoyancy pressure perturbation anomaly near the convective updrafts. The favorable horizontal PGF extended and strengthened the RTF inflow as a result. We could, therefore, expect a more effective hydrometeor recirculation process and more sustainable updrafts for these systems. Elsewhere, cold pool strengthening could be expected for these systems due to robust rainwater breakup-evaporation (Section 6.2). The horizontal expansion of the RIJ also caused the CIs to occur closer to the system edge.

The system RIJ was weaker and less horizontally expanded for experiments without rimed particle size sorting, which limited hydrometeor recirculation. The RIJs for these experiments
were weaker due to a combination of (a) decreased latent heat release aloft, and (b) rearward shift in rimed particle sedimentation location, which corresponded to the lack of preferential sedimentation in these experiments. From a vorticity balance perspective, the varied RIJ responses also fundamentally altered the system circulation: horizontal vorticity associated with the stronger RIJs could effectively negate the environmental vorticity and decreased the updraft tilt. Negative *vertical vorticity* was also generated near the updrafts through vorticity stretching and tilting, which could dynamically strengthen and extend the RIJ.

To conclude, the above analyses suggested that mid-level microphysics, especially the sedimentation characteristics of rimed particles, was the most significant factor that contributed to the cold pool variations and the overall differences in CI maintenance within the analyzed systems.

### 7.2. Implications of our findings

The varied responses of convective system characteristics when we only altered a small amount of model parameters related to rimed particle sedimentation were sobering and vividly illustrated the inherent model uncertainties when simplified drop-size distribution and parameterized fall speed relations were used. Our results suggested the effects of graupel sedimentation and size sorting on the accumulated precipitation and cold pool strength (which was correlated with near-surface wind potential) are substantial; They further illustrated the importance of the right end of graupel PSD spectrum (i.e. large drop) in adjusting the latent cooling rate below or near the melting level. This finding can have implications on some important scenarios of interest to the community, particularly convective systems’ response to aerosols. The importance of PSD pointed to the potential use
of dual-polarization radar simulators (e.g., Dawson et al. 2014; Snyder et al. 2017) in identifying and potentially correcting model microphysics biases. These results also hinted at potential improvement if bin microphysical schemes are used instead as hydrometeor drop-size distribution evolutions are explicitly simulated in these schemes.

Secondly, it was concluded in Chapter 5 that vertical moisture transport by sustained weak vertical updrafts in enhancing moisture depth was important in controlling the environmental precipitable water distributions. Based on this finding, we believe that increased assimilations of low-tropospheric moisture retrieval data by instruments such as WV-DIAL or AERI (Turner and Löhnert 2014) might be helpful in decreasing the forecast errors for the Great Plains nocturnal systems.

Finally, since our current study only focuses on one specific case that remained surface-based for a period of time during the nocturnal hours, the response analyzed in this study might be atypical and cautions will be needed in utilizing the findings presented in this study.

7.3. Future research paths

Future research work on this case will specifically focus on these general areas:

• The potential impacts of PECAN observational assimilation in correctly representing the observed boundary layer evolution.
• Severe near-surface wind potential when nocturnal stable boundary layers of varying strength are present.
• Perform statistical analyses on the PECAN nocturnal convective systems with in-situ microphysical observations collected by the NOAA P3 to study the hydrometeor drop-size
distributions within nocturnal MCSs and compare it to the IOP30 simulation results.

**Figures**

**Chapter 1**

![Diurnal cycle of mean precipitation amount](image)

**Fig.1.1.1** Diurnal cycle of mean precipitation amount between 1963 and 1993 during December-January-February (DJF; Upper left), March-April-May (MAM; Lower left), June-July-August (JJA; Upper right), and September-October-November (SON; Lower right), respectively. The arrow directions indicate the local time where maximum mean precipitation occur at a given location. Note that the JJA maximum precipitation occurrence time in the Great Plains is slightly after midnight. (Adopted from Fig.3 of Dai et al. 1999)
Fig.1.1.2 The tracks of large MCSs occurred between 1990 and 1992 in the Great Plains. Triangles represent the originating location for the convective systems. Circles with enclosed “x” represent the system centroids during their maximum extent. Circles with enclosed dots represent the system centroid during system dissipating stages. (Adapted from Fig.3a of Augustine and Caracena 1994).

Chapter 2
**Fig.2.1.1** RAP (a) 300-hPa and (b) 500-hPa analyses valid at 1200UTC, 14 July 2015. Red arrows represent the approximate locations of the mobile trough described in section 2.2. (c) RAP 850-hPa analysis valid at 0000UTC, 15 July 2015. Figure obtained from NCAR PECAN field catalog.

**Fig.2.1.2** (a) Water vapor satellite imagery at 2045UTC, 14 July 2015. Warm (Red-Maroon) colors represent regions with low atmospheric column vapor content. Colder colors (Grey-Blue) indicate high columnar vapor content. Figure obtained from Storm Prediction Center Event Archive. (b) RAP 500-hPa Vorticity Advection (shading) and Geopotential Height (contour) valid at 0000UTC, 15 July 2015.
Fig.2.1.3 Colorado topographic map.

Fig.2.2.1 Averaged 14 July 2015 2100 UTC sounding of the defined (a) NE CO region and (b) SE CO region (right). Location and extent of the two regions are shown in Fig.2.1.3.
Fig. 2.2.2 Storm Prediction Center (SPC) mesoanalysis on 0-6km bulk shear (kts) valid at 0000UTC, 15 July 2015. Figure obtained from Storm Prediction Center Event Archive.
Fig.2.2.3  MMM image archive radar reflectivity composite (dBZ) for (a) 0055UTC, (b) 0155UTC, (c) 0255UTC, (d) 0355UTC, (e) 0450UTC, (f) 0625UTC, (g) 0725UTC, (h) 0755UTC, and (i) 0825UTC, 15 July 2015. White arrows represent the radar fine line associated with cold pool propagation, red arrows represent the multiple fine lines associated with bore-soliton complex, yellow arrows represent main backbuilding convective initiation region. Cyan arrows denote the newly developed convective cells in MCS_N remnants.
Chapter 3

Fig. 3.1.1 Location of the deployed PECAN observation platforms during IOP30 (15 July 2015). The bold squares refer to the location of the fixed PISAs and soundings. The purple filled dots are the locations of the mobile PISAs and soundings. The pale blue stars represent the locations of available operational and research radar platforms. GOES-13 visible satellite imagery at 00:56UTC is overlaid to illustrate the approximate locations of the developing convective systems.

Fig. 3.2.1 Schematic diagram of the horizontal Arakawa C grid staggering utilized in WRF-ARW physical core. Figure obtained from Fig. 3.2 of Skamarock et al. (2008) doi:10.6342/NTU201900175
**Fig. 3.2.2** Domain configurations used in this study.

**Fig. 3.2.3** Simulated column-maximum radar reflectivity (dBZ) valid at 04UTC, 15 July 2015 using (a) NSSL two-moment microphysical scheme and (b) Morrison microphysical scheme.

<table>
<thead>
<tr>
<th>Grid Spacing</th>
<th>27km</th>
<th>9km</th>
<th>3km</th>
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<tbody>
<tr>
<td>Domain</td>
<td>155x155 grid points</td>
<td>271x271 grid points</td>
<td>445x334 grid points</td>
</tr>
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</table>

**Table 3.2.1** Domain specifications for all simulations discussed in this study.
**Chapter 4**

Fig. 4.1.1 (a) Pre-MCS$_N$ environmental $\theta_e$ vertical profile at 2359UTC (black), 0059UTC (brown), 0200UTC (blue), 0244UTC (green). (b) Same as (a), but for pre-MCS$_S$ profiles at 0007UTC (black), 0058UTC (brown), 0159UTC (blue) and 0259UTC (green).

Fig. 4.1.2 (a) Pre-MCS$_N$ environmental RH vertical profile at 2359UTC (black), 0059UTC (brown), 0200UTC (blue), 0244UTC (green). (b) Same as (a), but for pre-MCS$_S$ profiles at 0007UTC (black), 0058UTC (brown), 0159UTC (blue) and 0259UTC (green).
Fig.4.1.3 Surface temperature evolution relative to the cold pool passage time for FP5 (blue) and SPARC (orange). Note that the FP5 location was influenced by the northeast-propagating MCS$_S$ around 1-hr after the cold pool passage time, the temperature decrease afterwards was therefore not indicative of the intensification of the original MCS$_N$ cold pool.

Fig.4.1.4 Time (x axis) versus height (y axis) diagrams of observed pre-MCS$_N$ environmental variables from FP5 [Brewster, KS] sounding launches. (a) Potential temperature (gray contour, K) and potential temperature perturbation from 00UTC (shading, K). (b) Relative Humidity (shading, %) and water vapor mixing ratio (gray contours, g kg$^{-1}$).

Fig.4.1.5 As in Fig.4.1.4, but for pre-MCS$_S$ environmental variables from UW SPARC launches near Scott City, KS.
Fig.4.1.6 Pre-MCS$_N$ [FP5, Brewster, KS] vertical profiles of (a) CAPE (J kg$^{-1}$) at 2359UTC (pink line), 0059UTC (blue line), 0200UTC (green line), 0244UTC (black line) and 0357UTC (lemon chiffon line). (b) As in (a), but for CIN (J kg$^{-1}$) (c) As in (a), (b), but for $\Delta$LFC (km).

Fig.4.1.7 As in Fig.4.1.6, but for pre-MCS$_S$ launches at 0007UTC (red line), 0058UTC (blue line), 0159UTC (green line), 0259UTC (black line) and 0356UTC (lemon chiffon line).
Fig. 4.2.1 Vertical profiles of (a) sounding-derived horizontal divergence (gray lines) and vertical pressure-velocity (red lines) above triangle T1 centroid at 0200 UTC. (b) Same as (a), but above the triangle T2 centroid at 0400 UTC.

Fig. 4.2.2 (a) 0400 UTC NEXRAD radar mosaic of 4 km reflectivity (shading). The two triangles show the locations of triangle T1 (FP5-FP3-MP3; gray) and T2 (MP3-FP3-FP2; pale green). (b) Same as (a), but the triangles now show the locations of triangle T3 (MP3-FP2-FP6; gray) and T4 (MP2-FP4-FP6; pale green), respectively. Vectors represent the 0400 UTC RAP mean horizontal wind within the 700-850 hPa layer.
Fig.4.2.3 Comparison of sounding-derived vertical pressure-velocity profiles above the T3 centroid (red lines) and T4 centroid (blue lines) at 0400 UTC.

Fig.4.2.4 (a) RAP 2 m temperature (shading, °C) and 10 m wind field (vectors) at 0400 UTC. (b) Same as (a), but for the 2 m vapor mixing ratio (shading, g kg⁻¹) at 0400 UTC.
Fig. 4.2.5 (a) RAP 700-850 hPa mean temperature (shading, °C), wind field (vectors) and temperature advection (K hr⁻¹) at 0400 UTC. (b) Same as (a), but for the 700-850 hPa mean vapor mixing ratio (shading, g kg⁻¹). (c) Same as (a, b), but for the 700-850 hPa mean vertical pressure velocity (shading, Pa s⁻¹). (d) Same as (a, b, c), but for the precipitable water (shading, mm).
Fig. 4.2.6 (a) RAP 700-850 hPa mean horizontal moisture advection (shading, g kg\(^{-1}\) hr\(^{-1}\)) and wind field (vectors) at 0400 UTC. (b) Same as (a), but for the vertical moisture advection (shading; g kg\(^{-1}\) hr\(^{-1}\)) at 0400 UTC.

Fig. 4.2.7 (a) RAP 700-850 hPa mean frontogenesis (shading, K (100km\(^{-1}\)) hr\(^{-1}\)) and wind field (vectors) at 0400 UTC. (b) The vertical structure of frontogenesis (shading), vertical velocity (contours, m s\(^{-1}\)) along the cross section denoted by the red line in (a). Vectors denote the environmental wind paralleled to the cross section.
Fig. 4.2.8 (a) FP4 ceilometer backscatter cross section from 0230 UTC to 0600 UTC. (b) Time evolution of FP3 WV-DIAL relative backscatter during IOP30.
**Fig. 4.3.1** KGLD NEXRAD (Goodland, Kansas) 0.5° PPI observations at 0300 UTC. (a) Horizontal reflectivity \[Z_H \], (b) radial velocity, (c) same as (b), but for the 1.3° elevation angle. Red dot in the plot represents the radar location, black square represents the location of FP5, whereas the solid and dashed circle are the 50 km and 100 km range circle, respectively.

**Fig. 4.3.2** KGLD NEXRAD cross section of (a) \[Z_H \] and (b) radial velocity along the 270° azimuth at 0300 UTC. Grey dashed line represents the 0°C height (i.e. melting level).
Fig. 4.3.3 Same as Fig. 4.3.1, but for the SPoLKa $0.5^\circ$ PPI observations at 0535 UTC. (a) $Z_H$, (b) radial velocity. Red dot in the plot represents the radar location, black square represents the location of FP3, whereas purple line represents the $300^\circ$ azimuth angle used in Fig. 4.3.4 and Fig. 4.3.11.

Fig. 4.3.4 Same as Fig. 4.3.2, but for the SPoLKa RHI along the $300^\circ$ azimuth angle at 0535 UTC. (a) $Z_H$, (b) radial velocity.
Fig. 4.3.5 Same as Fig. 4.3.1, but for the KGLD NEXRAD 0.5° (a) differential reflectivity ($Z_{DR}$; dB), (b) correlation coefficient ($\rho_{HV}$), and (c) specific differential phase ($K_{DP}$; ° km$^{-1}$) at 0300 UTC. Red dot and black square represent the radar location and FP5 location, respectively.
**Fig. 4.3.6** Same as **Fig. 4.3.5**, but for the SPolKa 0.5° PPI observations at 0535 UTC. Red dot and black square represent radar location and FP3 location, respectively.

**Fig. 4.3.7** Interpolated (a) NEXRAD KGLD, and (b) SPolKa $Z_H$ at 0300 UTC and 0535 UTC, respectively. The black boxes denote the region where the average polarimetric variable profiles were derived.
Fig. 4.3.8 Spatially-averaged vertical profiles of (a) MCS\textsubscript{N} \(Z_H\) at 0300 UTC and (b) MCS\textsubscript{S} \(Z_H\) at 0535 UTC.

Fig. 4.3.9 Comparison of spatially-averaged (a) \(Z_{DR}\), (b) \(\rho_{HV}\), and (c) \(K_{DP}\) vertical profiles for MCS\textsubscript{N} (blue) and MCS\textsubscript{S} (red).
**Fig. 4.3.10** Ice habit diagram on the evolution of preferred crystal shape with ambient air temperature between 0°C and -70°C. (Adopted from Fig. 5 in Bailey and Hallett 2009)

**Fig. 4.3.11** Same as Fig. 4.3.4, but for (a) $Z_{DR}$ and (b) $K_{DP}$ at 0535 UTC. Black dashed line denotes the approximate location of the melting level.

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Chapter 5

Fig. 5.1.1 Simulated reflectivity (unit: dBZ) at (a) 0100UTC, (b) 0200UTC, (c) 0400UTC and (d) 0500UTC.

Fig. 5.2.1 (a) Simulated low-tropospheric vapor mixing ratio profiles interpolated to FP5 location (dashed lines) and SPARC location (transparent solid lines) during the pre-convective period. (b) Same as (a), but for potential temperature.

doi:10.6342/NTU201900175
Fig.5.2.2 Simulated Integrated Vapor Transport (IVT, unit: kg/ms) at (a) 0100UTC (shading) and 820hPa wind field (vector). (b) Same as (a), but for surface wind field (vector). (c,d) Same as (a,b), but at 0400UTC.
**Fig.5.2.3** WRF-simulated Integrated Water Vapor (IWV, unit: cm) at (a) 0100UTC (shading) and surface wind field (vector). (b) 0400UTC IWV (shading) and 820hPa wind field (vector).

**Fig.5.2.4** 0400UTC RAP analysis-derived (a) IVT (shading) and (b) IWV (shading). Vectors represent 820hPa wind field.
Fig. 5.2.5 WRF-simulated Integrated Moisture Convergence (IMC; unit: mm/hr) at (a) 0100UTC (shading) and surface wind field (vector). (b) 0400UTC IMC (shading) and 820hPa wind field (vector).

Fig. 5.2.6 (a) 750hPa $\omega$ (shading) for RAP analysis valid at 0400UTC (b) WRF-simulated 750hPa $\omega$ (shading) and wind field at the same pressure level (vector).
Chapter 6

Fig. 6.1.1 Schematic model on the microphysical impact of strong rear-inflow-jets (RIJs) to updraft strengthening near the melting level (Adopted from Fig. 6 of Siegel and van den Heever [2013]).

Fig. 6.1.2 (a) Terminal velocity–diameter relations for the three sensitivity tests in Group II. The red, grey, blue solid lines represent FERR, FULL and HDGR experiment, respectively. (b) Same as (a), but for HDGR (blue solid) and HAIL (blue dashed).
### Group I

<table>
<thead>
<tr>
<th>Experiment</th>
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<tr>
<td><strong>FULL</strong></td>
<td>Baseline experiment</td>
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<td><strong>GNSS</strong></td>
<td>Size-sorting disabled for graupel class</td>
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<tr>
<td><strong>RNSS</strong></td>
<td>Size-sorting disabled for rainwater class</td>
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<td><strong>GRNSS</strong></td>
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### Group II

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<th>Experiment</th>
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<td><strong>HDGR</strong></td>
<td>Same as FULL, but $\rho_g = 900\text{ g kg}^{-1}$</td>
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<tr>
<td><strong>FERR</strong></td>
<td>Same as FULL, but used the Ferrier (1994) graupel fall speed-diameter relationship.</td>
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### Group III

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<th>Experiment</th>
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<tr>
<td><strong>HAIL</strong></td>
<td>Same as FULL, but all larger frozen droplets are classified as hail instead.</td>
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#### Table 6.1.1 Descriptions of the sensitivity tests used in this study.

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<th>$d$</th>
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<td><strong>FULL</strong></td>
<td>$V(D) = \gamma c D^d$</td>
<td>$c=4\rho_g / (3C_0 \rho_a)^{-1}; C_0 = 0.8; \rho_g = 500\text{ g kg}^{-1}$</td>
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<tr>
<td><strong>FERR</strong></td>
<td>$V(D) = \gamma c D^d$</td>
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<td>0.37</td>
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<tr>
<td><strong>HDGR</strong></td>
<td>$V(D) = \gamma c D^d$</td>
<td>$c=4\rho_g / (3C_0 \rho_a)^{-1}; C_0 = 0.8; \rho_g = 900\text{ g kg}^{-1}$</td>
<td>0.5</td>
</tr>
<tr>
<td><strong>HAIL</strong></td>
<td>$V(D) = \gamma c D^d$</td>
<td>$c=4\rho_h / (3C_0 \rho_a)^{-1}; C_0 = 0.45; \rho_h = 900\text{ g kg}^{-1}$</td>
<td>0.5</td>
</tr>
</tbody>
</table>

#### Table 6.1.2 Summary of the different rimed particle terminal velocity-diameter relations used in this study.

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Fig. 6.2.1 (a) SPolKa observed 3 km CAPPI reflectivity at 0718UTC. Black line shows the cross section path of Fig. 6.2.2a. (b) FULL-simulated S-band reflectivity at 0400UTC.

Fig. 6.2.2 (a) Vertical structure of SPolKa reflectivity along the cross section shown in Fig. 6.2.1a. (b) Line-averaged cross section of FULL reflectivity at 0400UTC.

Fig. 6.2.3 (a) GNSS-simulated S-band reflectivity at 0400UTC. (b) RNSS-simulated S-band reflectivity at 0500UTC.
Fig. 6.2.4 (a) Line-averaged cross section of GNSS reflectivity at 0400UTC. (b) Line-averaged cross section of RNSS reflectivity at 0500UTC.

Fig. 6.2.5 (a) GRNSS-simulated S-band reflectivity at 0421UTC. (b) Line-averaged cross section of GRNSS reflectivity at 0421UTC.

Fig. 6.2.6 (a) HDGR-simulated S-band reflectivity at 0503UTC. (b) FERR-simulated S-band reflectivity at 0445UTC.
**Fig. 6.2.7** (a) Line-averaged cross section of HDGR reflectivity at 0509UTC. (b) Line-averaged cross section of FERR reflectivity at 0445UTC.

**Fig. 6.2.8** (a) HAIL-simulated S-band reflectivity at 0512UTC. (b) Line-averaged cross section of HAIL reflectivity at 0512UTC.
Fig.6.2.9 (a) Line-averaged cross section of FULL graupel mixing ratio (g kg\(^{-1}\), shading), cloud water mixing ratio (black contours), rainwater mixing ratio (magenta contours), snow mixing ratio (blue contours), and ice mixing ratio (red contours) at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC. (c) Same as (a-b), but for RNSS at 0500UTC. (d) Same as (a-c), but for GRNSS at 0421UTC.

Fig.6.2.10 Same as Fig.6.2.9, but for (a) HDGR at 0509UTC and (b) FERR at 0445UTC, respectively.

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Fig. 6.2.11 Same as Fig. 6.2.9-10, but for HAIL at 0512 UTC.

(a) 56.6 mm  
(b) 44.7 mm  
(c) 62.4 mm  
(d) 37.8 mm
Fig 6.2.12 Distribution of 4-hr accumulated precipitation (0300UTC-0700UTC; Shading) for (a) FULL, (b) GNSS, (c) RNSS, (d) GRNSS, (e) HDGR, (f) FERR, and (g) HAIL, respectively. The numerical values shown in the bottom-right part of these plots represent the maximum 4-hr accumulated precipitation grid-point values for simulated MCSs.
**Fig.6.2.13** Averaged mean-volume rainwater diameter profiles within the FULL (red), RNSS (black), and HDGR (blue) MCSs convective regions, defined here as the region with reflectivity values greater than 30 dBZ.

**Fig.6.2.14** (a) Line-averaged cross section of HAIL rimed particle number concentration (m^-3, shading), reflectivity (colored contours), and melting level (T=0 °C; brown line) at 0521UTC. (b) Same as (a), but for FULL at 0400UTC. (c) Same as (a-b), but for RNSS at 0500UTC.

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Fig.6.2.15 Air Temperature (°C, shading), wind field (vectors), and reflectivity (contour) at the lowest model model for (a) FULL at 0400UTC, (b) GNSS at 0400UTC, (c) RNSS at 0500UTC, (d) GRNSS at 0421UTC, (e) HDGR at 0509UTC, (f) FERR at 0354UTC, and (g) HAIL at 0521UTC.

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Fig. 6.2.16 (a) Averaged latent cooling rate profiles within the FULL (red), RNSS (black), GNSS (blue), and GRNSS (green) MCSs convective regions, defined here as the region with reflectivity values greater than 30 dBZ. (b) Same as (a), but for relative humidity. (c) Same as (a-b), but for rainwater number concentration. (d) Same as (a-c), but for downdrafts ($w<0$ m s$^{-1}$). (e) Same as (a-d), but for graupel number concentration.
Fig. 6.3.1 Box-and-whisker plots of updraft strength for (a) FULL (brown boxes) and GNSS (gray boxes) during the denoted time period. (b) FULL (brown boxes) and RNSS (gray boxes) during the MCS developing, mature, weakening phases. There is a 1-hour time difference between FULL and RNSS to account for the time bias discussed in the text. (c) Same as (a), but for GRNSS (gray boxes).
Fig. 6.3.2 Strong updrafts ($w > 2 \text{ m s}^{-1}$) CFADs during the (a) system developing phase (i.e. 0230-0330 UTC), (b) system mature phase (i.e. 0330-0430 UTC), and (c) system weakening phase (i.e. 0430-0530 UTC). This plot compares the 0.135 frequency contours for the FULL (red solid) and the GNSS (yellow dashed). White dashed lines represent the approximate height for the 0 °C level. For reference, the FULL CFAD is shown as colored contours.
Fig. 6.3.3 Same as Fig. 6.3.2, but for the FULL (red solid) and the RNSS (blue dashed) during the (a) developing phase (i.e. 0230-0340UTC FULL; 0330-0430UTC RNSS), (b) mature phase (i.e. 0330-0430UTC FULL; 0430-0530UTC RNSS), and (c) weakening phase (i.e. 0430-0530UTC FULL; 0530-0630UTC RNSS, respectively.)
Fig. 6.3.4 Reflectivity (shading) and wind field (vector) at 4 km for Group I simulations. (a) represents FULL at 0400UTC, (b) represents GNSS at 0400UTC, (c) represents RNSS at 0500UTC, (d) represents GRNSS at 0421UTC.
Fig.6.3.5  (a) Line-averaged cross section of FULL line-normal storm-relative inflow wind (shading) at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC. (c) Same as (a-b), but for RNSS at 0500UTC.

Fig.6.3.6  Same as Fig.6.3.5, but the analysis time is set to (a) 0245UTC for FULL, (b) 0245UTC for GNSS, and (c) 0354UTC for RNSS, respectively.
Fig. 6.3.7 (a) Line-averaged cross section of FULL latent cooling rate (K s⁻¹; shading), graupel mixing ratio (g kg⁻¹, contours) and storm-relative wind (vectors) at 0245UTC. (b) Same as (a), but for GNSS at 0245UTC. (c) Same as (a-b), but for RNSS at 0354UTC.
Fig. 6.3.8 Same as Fig. 6.3.7, but for the relative humidity (%, shading), graupel mixing ratio (contours) and storm-relative inflow wind (vectors) near the convective region.

Fig. 6.3.9 (a) Average profile of RH_{water} evolution under different air temperature conditions for FULL (black), GNSS (blue) and RNSS (red) within the region with substantial latent heating (<=-0.0015 K s^{-1}) in the vicinity of convective updrafts. (b) Same as (a), but for RH_{ice}. 

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Fig.6.3.10 (a) Average buoyancy profile within the region used in Fig.6.3.9. Black line represents FULL, while gray line represents GNSS. (b) Same as (a), but for the vertical buoyancy

Fig.6.3.11 (a) Line-averaged cross section of FULL buoyancy (m s$^{-2}$; shading), 20 and 30 dBZ outline (thick gray contours) and storm-relative wind (vectors) at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC. (c) Same as (a-b), but for RNSS at 0500UTC. (d) Same as (a-c), but for GRNSS at 0421UTC.

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Fig. 6.3.12  (a) Line-averaged cross section of FULL latent heating rate \([K \text{ (5 min)}^{-1}]\); red shading), 20 and 30 dBZ outline (thick gray contours), storm-relative wind (vectors), and latent cooling rate (blue shading) at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC. (c) Same as (a-b), but for RNSS at 0500UTC. (d) Same as (a-c), but for GRNSS at 0421UTC.
Fig.6.3.13 (a) Line-averaged cross section of FULL mean-volume graupel diameter (mm; shading), graupel mixing ratio (thick gray contours), storm-relative wind (vectors), and graupel number concentration near convective region at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC.

Fig.6.3.14 PDFs (Probability Density Functions) for mean-volume graupel diameter above melting level (T=0 °C) for the four Group I simulations: FULL at 0400UTC (magenta), GNSS at 0400UTC (red), RNSS at 0500UTC (green), and GRNSS at 0421UTC (blue).

Fig.6.3.15 Box-and-whisker plots of graupel number concentration (m⁻³) for (a) FULL (brown boxes) and GNSS (gray boxes) during the denoted time period.
Fig. 6.3.16 Joint distribution of graupel number concentration and air temperature during system mature phase. (a) compares the kernel density 0.025 contours for the FULL (red) and GNSS (yellow), whereas (b) compares the FULL (red) and the RNSS (yellow). In both plots, the FULL distribution was shaded for reference.

Fig. 6.3.17 (a) Line-averaged cross section of FULL latent cooling rate [K (5 min)$^{-1}$; shading], vapor increase tendency (red contours), and storm-relative wind (vectors) near convective region at 0400UTC. (b) Same as (a), but for GNSS at 0400UTC. (c) Same as (a), but vapor decrease tendency (red contours).
Fig. 6.4.1 Same as Fig. 6.3.2, but for the HDGR experiment (light orange solid).

Fig. 6.4.2 Same as Fig. 6.3.7, but for the (a) FULL experiment at 0245 UTC and (b) HDGR experiment at 0345 UTC (right column).
**Fig. 6.4.3** Same as Fig. 6.3.2 and Fig. 6.4.1, but for the FERR experiment (light green solid). (d) is similar to (c), but the 0.01 frequency contours are shown.

**Fig. 6.4.4** Storm-relative inflow (m s\(^{-1}\)) during the mature phase for (a) the FULL experiment at 0509 UTC and (b) HDGR experiment at 0445 UTC.
Fig. 6.4.5 Same as Fig. 6.3.15, but for the (a) HDGR experiment at 0509 UTC and (b) FERR experiment at 0445 UTC.

Fig. 6.4.6 Same as Fig. 6.3.14, but for the (a) HDGR experiment at 0509 UTC, (b) FERR experiment at 0445 UTC, and (c) FULL at 0400 UTC, respectively. Note that (b) was horizontally zoomed in due to the narrowness of the system.
Fig. 6.4.7 Averaged (a) buoyancy and (b) buoyancy gradient profiles within the graupel sedimentation region (definition same as that used in Fig. 6.3.10 and Fig. 6.3.11), but for the FULL experiment (black lines) and FERR experiment (gray lines). Brown dashed line represents the melting level (0 °C line).
**Fig.6.4.8** Same as **Fig.6.3.2**, but for the HAIL experiment (cyan solid).

**Fig.6.4.9** Same as **Fig.6.4.6**, but for the (a) HAIL experiment at 0512 UTC and (b) FULL experiment at 0400 UTC. The HAIL result is zoomed horizontally, similar to **Fig.6.4.6(b)**.

**Fig.6.4.10** Same as **Fig.6.4.6**, but for the HAIL experiment at 0512 UTC.
Fig.6.4.11 Same as Fig.6.4.7, but for the FERR experiment (black lines) and HAIL experiment (gray lines).

Fig.6.4.12 (a) Same as Fig.6.3.19, but for the HAIL experiment at 0512 UTC. (b) Same as Fig.6.3.19c, but for the HAIL experiment at 0512 UTC. (c) Same as (a), but for the cloud drop decrease tendency (shading; g kg\(^{-1}\) hr\(^{-1}\)).
**Fig. 6.5.1** (a) Line-averaged cross section of FULL $P_b$’ (shading, Pa) at 0400 UTC. Black dashed line denoted the approximate location of the convective updraft axis. (b) Same as (a), but for the GNSS at 0400 UTC. Gray dashed line denoted the updraft axis, while the black dashed line represented the descending inflow layer.

**Fig. 6.5.2** (a) Same as **Fig. 6.5.1(a)**, but for the line-averaged, system-normal component of the wind speed change rate induced by the PGF$_P$ for the FULL (shading; m s$^{-2}$) at 0400 UTC. (b) Same as **Fig. 6.5.1(b)**, but for the line-averaged, GNSS system-normal component of wind speed change rate induced by PGF$_P$ at 0400 UTC.
Fig. 6.5.3 (a) Same as Fig. 6.5.1(a) and Fig. 6.5.2(a), but for the dynamical pressure perturbation (shading; Pa) at 0400 UTC. (b) Same as Fig. 6.5.1(b) and Fig. 6.5.2(b), but for the dynamical pressure perturbation (shading; Pa) at 0400 UTC.

Fig. 6.5.4 (a) Horizontal map of the FULL system at 2.5 km AGL, valid at 0400 UTC. Gray shading denotes regions with reflectivity higher than 35 dBZ, black contours represent vertical vorticity (plotted every 0.5 x 10^{-3} s^{-1}), while red vectors denote the horizontal vorticity direction near the FULL system (the vectors were divided by 10^{-3} for plotting purposes). The solid gray line represents the cross-section utilized in Fig. 6.5.7. (b) Same as (a), but for the GNSS system at 2.5 km AGL, valid at 0400 UTC. Gray solid line in this plot represents the cross-section utilized in Fig. 6.5.8.
Fig.6.5.5 (a) 2.5 km AGL horizontal reflectivity (shading) for the FULL system and horizontal wind (vectors) at 0400 UTC. (b) Same as (a), but for the GNSS at 0400 UTC.

Fig.6.5.6 (a) Same as Fig.6.5.4(a), but for vertical velocity at 2.5 km AGL (solid contours: upward motion, dashed contours: downward motion; plotted every 0.5 m s\(^{-1}\)). (b) Same as (a), but for the GNSS system at 0400 UTC. (c) Same as (a), but for vorticity tilting and stretching at 2.5 km AGL (plotted every 0.5 x 10\(^{-6}\) s\(^{-2}\)). (d) Same as (c), but for the GNSS system at 0400 UTC.
Fig. 6.5.7 Vertical cross section of FULL system along the line AA’ (location shown in Fig. 6.5.4a), valid at 0400 UTC. The variables depicted include: (a) Vertical vorticity (shading; s$^{-1}$ x 10$^{-3}$), vertical velocity (black contours; m s$^{-1}$) and vorticity vectors that paralleled AA’.
(b) Summation of vorticity tilting and stretching (shading; s$^{-2}$ x 10$^{-6}$), vertical vorticity (black contours; s$^{-1}$ x 10$^{-3}$). (c) Horizontal vorticity (shading; s$^{-1}$ x 10$^{-3}$), wind field that paralleled AA’ (vector; m s$^{-1}$).

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Fig. 6.5.8 Same as Fig. 6.5.7, but for the vertical cross sections of GNSS system along BB' (location shown in Fig. 6.5.4b).
Appendix 1 - Data processing and quality control

(a) Radars

Three quality control methods are adopted in this study: (i) non-meteorological signal filtering; (ii) calibration of ZDR; and (iii) doppler velocity unfolding. A simple dual-polarization variable-based multistep filter was applied to the raw radar data to eliminate non-meteorological signals. First, range gates with $\rho_{hv}<0.8$ and $Z_{DR}>6$ dB were removed to filter out objects with high anisotropy. Secondly, range gates with spectrum width $>8.0$ were also removed. Thirdly, a low reflectivity threshold of 7 dBZ was set to exclude Bragg scattering. Finally, small and isolate objects were masked out by a despeckle algorithm available in the open-source Py-ART package. An example is shown in Fig.A2.1.1a,b.

Systematic bias in $Z_{DR}$ observations that occur either randomly or due to poor radar calibration could cause significant errors in radar-based quantitative precipitation estimation (QPE) and hydrometeor classification. These errors are often substantial for NEXRAD radars (Homeyer and Kumjian 2015). $Z_{DR}$ calibration was performed in real-time for the SPolKa during PECAN as vertically-pointed radar scans were available. For the NEXRAD radars, an alternative $Z_{DR}$ calibration approach was adopted: $Z_{DR}$ probability density functions (PDFs) was derived for the dry snow region, the $Z_{DR}$ bias can subsequently be determined by the difference between the PDF peak and zero\(^{23}\). Through this procedure, the systematic bias for KGLD was determined to be $\sim 0.26$ dB (Fig.A2.1.2a). Finally, regions with unphysical velocity texture or gradient were unfolded by a region-based unfolding algorithm. Fig.A2.1.3a,b illustrate an example of this procedure.

\(^{23}\) This criteria is adopted by dry snow region tends to have near zero $Z_{DR}$. 

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Fig.A1.1.1 KGLD 0.5° reflectivity (a) before applying quality-control procedure, (b) after applying quality-control procedure.

Fig.A1.1.2 Probability density functions (PDFs) for (a) WSR-88D radar at Goodland, KS (KGLD) at 02:23UTC [MCSN]. (b) SPoIka radar at 06:30UTC [MCS].

Fig.A1.1.3 SPoIka 0.5° radial velocity at 06:10UTC (a) before velocity unfolding, (b) after unfolding. The red dot refers to the location of SPoIka. The two filled squares represent the locations of the fixed FP3 disdrometer (black) and mobile disdrometer (blue) respectively.
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