Sensitivity of a Simulated Midlatitude Squall Line to Parameterization of Raindrop Breakup

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ABSTRACT

This paper describes idealized simulations of a squall line observed on 20 June 2007, in central Oklahoma. Results are compared with measurements from dual-polarization radar and surface disdrometer. The baseline model configuration qualitatively reproduces key storm features, but underpredicts precipitation rates and generally overpredicts median volume raindrop diameter. The sensitivity of model simulations to parameterization of raindrop breakup is tested under different low-level (0–2.5 km) environmental vertical wind shears. Storm characteristics exhibit considerable sensitivity to the parameterization of breakup, especially for moderate (0.0048 s$^{-1}$) shear. Simulations with more efficient breakup tend to have higher domain-mean precipitation rates under both moderate and higher (0.0064 s$^{-1}$) shear, despite the smaller mean drop size and hence lower mass-weighted fall speed and higher evaporation rate for a given rainwater content. In these runs, higher evaporation leads to stronger cold pools, faster propagation, larger storm size, greater updraft mass flux (but weaker convective updrafts at mid- and upper levels), and greater total condensation that compensates for the increased evaporation to give more surface precipitation. The impact of drop breakup on mass-weighted fall speed is also important and leads to a nonmonotonic response of storm characteristics (surface precipitation, cold pool strength, etc.) to changes in breakup efficiency under moderate wind shear. In contrast, the response is generally monotonic at higher wind shear. Interactions between drop breakup, convective dynamics, cold pool intensity, and low-level environmental wind shear are also described in the context of “Rotunno–Klemp–Weisman (RKW) theory,” which addresses how density currents evolve in sheared environments.

1. Introduction

Squall lines are linear systems of organized deep convection, or mesoscale convective systems (MCSs). They are common in both midlatitude and tropical environments and have been extensively studied (e.g., Fujita 1955; Zipser 1969; Ogura and Chen 1977; LeMone et al. 1984; Houze 1989; Biggerstaff and Houze 1991, 1993). These studies have suggested several common morphological features described by the conceptual model of Biggerstaff and Houze (1991, see Fig. 18 therein). Mature squall lines typically contain an upshear-tilted, multicellular convective region with heavy precipitation and active updraft cell generation along the gust front, a transition zone of lighter precipitation and a low-level radar reflectivity minimum between the convective and stratiform regions, followed by a region of moderate precipitation in the trailing stratiform region.

Cloud-system-resolving models (CSRM)s utilizing a horizontal grid spacing of order 1 km are well suited for simulating squall lines and other MCSs because they can explicitly resolve the mesoscale and deep convective-scale dynamics. Understanding the behavior of such simulations is important since regional numerical weather prediction (NWP) models are now commonly run at CSRM scale for convective-scale forecasting (e.g., Kain et al. 2008; Lean et al. 2008; Weisman et al. 2008). Furthermore, CSRMs have been employed as smaller-scale models embedded within larger-scale general circulation models (GCMs; i.e., the multiscale modeling framework or “superparameterization,” primarily to improve the representation of deep convection; e.g., Grabowski 2001; Randall et al. 2003; Tao et al. 2009).

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The parameterization of cloud microphysics remains a major challenge for simulating squall lines using CSRMs. Microphysical processes directly impact buoyancy, and hence convection, through condensate loading and latent heating or cooling due to phase changes. Several studies have investigated the sensitivity of squall lines to representation of microphysics using bulk microphysics schemes (e.g., Fovell and Ogura 1988; McCumber et al. 1991; Tao et al. 1995; Ferrier et al. 1995; Liu et al. 1997; Morrison et al. 2009; Bryan and Morrison 2012; van Weverberg et al. 2012). Parameterization of rain microphysics has been identified as a critical aspect given the impact of rain evaporation on cold pool evolution (Ferrier et al. 1995; Morrison et al. 2009; Bryan and Morrison 2012; van Weverberg et al. 2012). Morrison et al. (2009), Bryan and Morrison (2012), and van Weverberg et al. (2012) investigated sensitivity to using either a one-moment (predicting mass mixing ratio only) or two-moment scheme (predicting both mass and number mixing ratio) representation of rain, with all other microphysical processes treated identically. Differences in the rain drop size distributions (DSDs) between the one- and two-moment schemes led to reduced evaporation, enhanced surface precipitation in the trailing stratiform region, and weaker cold pools (especially well behind the surface gust front) using the two-moment schemes, with an overall improvement compared to observations (Bryan and Morrison 2012). Similar impacts using one-moment versus two-moment schemes have also been described for other types of deep convection (Milbrandt and Yau 2006; Mansell and Wicker 2008; Luo et al. 2010; Dawson et al. 2010), sometimes, but not always, leading to an improved comparison with observations when utilizing two-moment schemes.

Despite sensitivity to representation of rain DSDs, there has been little testing of specific processes, such as raindrop breakup, that influence the DSDs in squall-line simulations using two-moment schemes. An exception is van Weverberg et al. (2012), who found that differing treatments of breakup led to large differences in maximum accumulated precipitation using two different two-moment schemes. More systematic testing of sensitivity to drop breakup for a supercell storm was performed by Morrison and Milbrandt (2011, hereafter MM11). They found that evaporation and hence cold pool characteristics were highly sensitive to the parameterization of breakup; minimum perturbation potential temperature varied by up to 7 K after 2 h of integration using various drop breakup formulations proposed in the literature. In these simulations, more efficient breakup parameterizations led to reduced surface precipitation and stronger cold pools (see Table 2 in MM11). These impacts are consistent with a decrease in mean drop size as breakup efficiency is increased, leading to reduced fall speed and increased evaporation. MM11 concluded that constraining the parameterization of raindrop breakup is needed to reduce uncertainty of two-moment microphysics schemes, which is important since these schemes are expected to be widely used in NWP and climate models in the coming years. In this study, we extend the work of MM11 and van Weverberg et al. (2012) to test systematically the sensitivity of a simulated squall line to drop breakup.

Environmental wind shear plays a critical role in the organization and structure of squall lines (e.g., Thorpe et al. 1982; Nicholls et al. 1988; Weisman et al. 1988; Fovell and Ogura 1989; Szeto and Cho 1994; Robe and Emanuel 2001; Weisman and Rotunno 2004; James et al. 2005; Bryan et al. 2006; Takemi 2007). Therefore we test sensitivity of the simulated squall line to drop breakup over a range of different low-level vertical shears. Interactions between drop breakup, cold pools, and storm dynamics under varying strength of low-level shear can also be viewed in the context of Rotunno–Klemp–Weisman (RKW) theory, which explains how density currents evolve in a sheared environment (Rotunno et al. 1988; Weisman and Rotunno 2004). This theory postulates that maximum system intensity is attained when the vorticity generated by cold pools is in approximate balance with the low-level vertical environmental shear, leading to the deepest lifting of environmental air. Hence, changes in cold pool intensity or shear that move the system away from optimal balance may result in weakening of the system. The simulations described in this paper provide a backdrop for analysis of interactions between cold pool evolution and low-level shear and hence assessment of RKW theory.

The overall goals of this study are 1) to develop a well-characterized midlatitude squall-line case study by synthesizing a suite of observations, 2) to assess model performance against these observations, and 3) to quantify sensitivity of the model to the treatment of raindrop breakup under different low-level environmental wind shears, with interactions between microphysics, cold pool evolution, and storm dynamics analyzed in the context of RKW theory. In this paper, the impacts of raindrop breakup on three-dimensional CSRSM simulations of a squall line are described. While this study focuses largely on sensitivity of the simulated storm to raindrop breakup, the comparison with observations demonstrates the

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1 Note that since breakup impacts number concentration but not bulk mass of rain, it is parameterized in two-moment but not one-moment schemes.
model’s ability to reproduce overall storm features and provides observational constraints on the simulated rain DSDs. The analysis focuses on several key squall-line properties, including surface precipitation, cold pool characteristics, radar reflectivity, and dynamical structure, with an emphasis on interactions between the microphysics, thermodynamics, and dynamics, and how these interactions respond to different parameterizations of breakup. The combination of disdrometer, radar, and other measurements for this case affords an opportunity to address this issue by providing robust observational constraints on the modeled DSDs.

2. Case description and observations

The 20 June 2007 squall line in central Oklahoma was part of a large MCS (Fig. 1) that resulted from three separate convective systems that formed during the afternoon of 19 June 2007 and merged together later that evening. The first storms developed by 1700 UTC 19 June in northwestern Kansas along a stationary front. This cluster of storms moved southeastward toward Oklahoma and merged with another set of storms that formed by 2100 UTC in northern Oklahoma, forming the portion of the squall line that passed over the Norman, Oklahoma (KOUN), dual-polarization radar domain (Fig. 1) that is the focus of the present study. This storm was the subject of a previous study on electrification by Lang et al. (2010). Southeasterly low-level flow had set up on 19 June, advecting moisture northward from the Gulf of Mexico into the south-central plains to the east of a dryline across eastern New Mexico. Extremely moist and unstable conditions prevailed over Oklahoma and Texas, with surface dew point temperatures exceeding 25°C and a convective available potential energy (CAPE) of 4113 J kg\(^{-1}\) shown by the sounding at 0000 UTC 20 June in Fig. 2 (calculated using 110-hPa mixed layer parcels).

The observational infrastructure in central Oklahoma for this case included two two-dimensional video disdrometers (2DVDs; Kruger and Krajewski 2002), a dual-polarization S-band Weather Surveillance Radar-1988 Doppler (WSR-88D) radar at KOUN, and a high-resolution network of surface observations from the Oklahoma Mesonet. The Oklahoma Mesonet (Brock et al. 1995; McPherson et al. 2007) provided surface measurements every 5 min, such as temperature, humidity, wind, and precipitation, at more than 100 sites in the state of Oklahoma. The 2DVDs were deployed within the KOUN radar domain during the 2007 wet season: one operated and owned by NCAR (17 km to the east of KOUN) and the other by the University of Oklahoma (OU; 65 km to the southwest of KOUN). The radar, mesonet, and disdrometer measurements were used to derive rain-rate and DSD parameters for
comparison with the model. A more detailed description of the methodology for processing and analyzing these measurements is provided in the appendix.

3. Model description and setup

a. Model description

Simulations were conducted using the compressible, nonhydrostatic Advanced Research Weather Research and Forecasting model (WRF) version 3.1 (Skamarock et al. 2008). The governing equations are solved using a time-split integration with third-order Runge–Kutta scheme. Horizontal and vertical turbulent diffusion are calculated using a 1.5-order TKE scheme (Skamarock et al. 2008). Horizontal and vertical advection is calculated using positive definite fifth- and third-order discretization schemes, respectively. The upper and lower boundaries are free slip with zero vertical velocity. Surface fluxes are set to zero and radiative transfer has been neglected for simplicity. Given that the focus of this paper is on treatment of microphysics, its parameterization is described in more detail below.

b. Description of the microphysics scheme

The two-moment bulk cloud microphysics scheme used here predicts mass and number mixing ratios of five hydrometeor species: cloud droplets, cloud ice, snow, rain, and hail. This scheme is based on the parameterization of Morrison et al. (2005), and subsequently implemented, with modifications, into WRF (Morrison et al. 2009). The hydrometeor size distributions $N(D)$ are treated using gamma functions such as

2 The Morrison scheme available in WRF has a switch to represent the rimed ice category as either graupel or hail. In this study, all simulations utilized settings representing rimed ice as hail.
where \( D \) is the particle diameter; and \( N_0, \mu, \) and \( \lambda \) are the intercept, shape, and slope parameters, respectively. For all precipitation species as well as cloud ice, we assume that \( \mu = 0 \) (i.e., an inverse exponential distribution). For cloud droplets, \( \mu \) is a function of the predicted droplet number concentration following the observations of Martin et al. (1994). Since the scheme is two moment, \( N_0 \) and \( \lambda \) are free parameters that are determined from the predicted mass and number mixing ratios for each species:

\[
\lambda = \left[ \frac{c \Gamma(\mu + 4)}{\Gamma(\mu + 1)} \right]^{1/d},
\]

\[
N_0 = \frac{\nu A \mu + 1}{\Gamma(\mu + 1)},
\]

where \( \Gamma \) is the Euler gamma function and the parameters \( c \) and \( d \) are given by the assumed power-law mass–diameter \( (m-D) \) relationship of the hydrometeors for each species, where \( m = cD^d \). Here, all particles are assumed to be spheres for simplicity, with a bulk particle density for the various ice species following Reisner et al. (1998), except for hail, for which the bulk density is equal to 900 kg m\(^{-3}\) following Lin et al. (1983). Other details of the parameterization, including formulations for the various microphysical process rates, are given by Morrison et al. (2009) and references therein. For simplicity, here we have assumed a constant cloud droplet concentration of 250 cm\(^{-3}\). A description of the calculation of radar reflectivity from simulated hydrometeor fields is given by Morrison et al. (2009).

The parameterization of raindrop breakup, which is the focus of this paper, follows from a modified version of the Verlinde and Cotton (1993, hereafter VC93) parameterization in the baseline model configuration (BASE).\(^3\) As in most other drop breakup schemes, VC93 represents collisional drop breakup by modifying the bulk collection efficiency for rain self-collection \( E_c \). Breakup leads to a decrease of \( E_c \) as a function of mean drop size \( D_r \) (equal to \( 1/\lambda \) for the assumed exponential DSD). Negative values of \( E_c \) can occur when \( D_r \) is large, implying a net source of drop number from the combined effects of drop breakup and self-collection. The formulation for \( E_c \) in the baseline scheme is given by

\[
E_c = 1, \quad D_r < D_{th},
\]

\[
E_c = 2 - \exp[2300(D_r - D_{th})], \quad D_r \geq D_{th},
\]

where \( D_{th} \) is the threshold drop size at which breakup occurs. For the baseline model configuration, \( D_{th} = 300 \) \( \mu \)m. Sensitivity to drop breakup was tested using different formulations described in the literature: Ziegler [1985, hereafter Z85, his Eq. (18)] and Seifert [2008, hereafter S08, his Eq. (A13)]. Specific parameter settings in these formulations are fairly arbitrary, although S08 showed their scheme produced reasonable results (in conjunction with parameterized rain evaporation and spectral shape parameter) in a one-dimensional rainshaft model when compared to a detailed bin microphysics model. Nonetheless, there is considerable uncertainty in these formulations and they produce large differences in \( E_c \) as a function of \( D_r \) (Fig. 3). A key difference between these formulations is that S08 and Z85 give a weaker reduction of \( E_c \) with increasing \( D_r \) compared to BASE; \( E_c \) values are greater than zero for Z85 for all values of \( D_r \) and hence Z85 has less efficient breakup overall. Additional sensitivity tests were also performed using the modified VC93 scheme as in BASE, but with \( D_{th} \) varied between 105 and 510 \( \mu \)m (BREAK1, BREAK2, and BREAK3). These tests allow us to isolate changes to a single parameter in the breakup scheme and more systematically investigate sensitivity to the efficiency of drop breakup. While values of \( D_{th} \) of a few hundred \( \mu \)m are small compared to sizes of large drops (greater than \( \sim 0.8 \) mm) that were observed to undergo breakup after colliding with smaller drops (Barros et al. 2008), it should be kept in mind that \( D_{th} \) represents a number-weighted mean drop size; the DSDs extend to drop sizes larger than \( D_{th} \). As shown in Fig. 3, BREAK1 gives more efficient breakup (lower \( E_c \)) relative to BASE, S08, or Z85. BREAK2 and BREAK3 produce less

\(^3\) Drop breakup using the modified VC93 formulation is included in the Morrison scheme in WRFV3.2 and subsequent versions. Breakup in previous versions was treated implicitly by limiting rain \( \lambda \) to a minimum value of 20 cm\(^{-1}\).
Table 1. List of the main simulations described in the text (shear indicates the specified 0–2.5-km environmental vertical wind shear).

<table>
<thead>
<tr>
<th>Name</th>
<th>Description</th>
<th>Shear (s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BASE</td>
<td>Baseline model configuration (modified VC93 breakup parameterization, $D_{th} = 300$ μm)</td>
<td>0.048</td>
</tr>
<tr>
<td>S08</td>
<td>Breakup parameterization of S08</td>
<td>0.048</td>
</tr>
<tr>
<td>Z85</td>
<td>Breakup parameterization of Z85</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK1</td>
<td>Modified VC93 breakup parameterization, $D_{th} = 105$ μm</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK2</td>
<td>Modified VC93 breakup parameterization, $D_{th} = 405$ μm</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK3</td>
<td>Modified VC93 breakup parameterization, $D_{th} = 510$ μm</td>
<td>0.048</td>
</tr>
<tr>
<td>BASEF</td>
<td>As in BASE, but for $V_m$ and $V_N$ independent of mean drop size</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK1F</td>
<td>As in BREAK1, but for $V_m$ and $V_N$ independent of mean drop size</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK2F</td>
<td>As in BREAK2, but for $V_m$ and $V_N$ independent of mean drop size</td>
<td>0.048</td>
</tr>
<tr>
<td>BREAK3F</td>
<td>As in BREAK3, but for $V_m$ and $V_N$ independent of mean drop size</td>
<td>0.048</td>
</tr>
<tr>
<td>BASEH</td>
<td>As in BASE, but for higher shear</td>
<td>0.064</td>
</tr>
<tr>
<td>BREAK1H</td>
<td>As in BREAK1, but for higher shear</td>
<td>0.064</td>
</tr>
<tr>
<td>BREAK2H</td>
<td>As in BREAK2, but for higher shear</td>
<td>0.064</td>
</tr>
<tr>
<td>BREAK3H</td>
<td>As in BREAK3, but for higher shear</td>
<td>0.064</td>
</tr>
</tbody>
</table>

efficient breakup (greater $E_c$) than S08 or Z85 at relatively small $D_n$, but more efficient breakup for $D_r > 0.6$–0.8 mm. To separate the impact of drop breakup on rain sedimentation from that on other processes, additional simulations were performed with the rain mass- and number-weighted fall speeds independent of mean drop size and hence breakup (see section 4c). A summary of the main simulations described herein is given in Table 1.

c. Experimental design

We use an idealized model setup similar to Bryan and Morrison (2012). Given the length of the observed squall line (several hundred kilometers; see Fig. 1), only a portion is simulated using periodic lateral boundaries in the along-line ($y$) direction. Open boundary conditions are used in the across-line ($x$) direction. Simulations are integrated for 8 h. The domain size is 612 km × 128 km in the horizontal with the model lid at 20 km. A Rayleigh damper with damping coefficient of 0.003 s$^{-1}$ is used in the upper 5 km to damp waves in the stratosphere. Horizontal and vertical grid spacings are 1 km and approximately 250 m, respectively. Previous studies have documented sensitivity of squall-line simulations to grid spacing (Bryan et al. 2003; Bryan and Morrison 2012). While Bryan and Morrison (2012) found some differences in quantities like surface precipitation and updraft mass fluxes as the horizontal grid spacing was decreased from 1 km to 250 m, these runs were qualitatively similar and microphysical sensitivities were relatively unaffected by grid spacing. Thus, we acknowledge model resolution as a potential source of uncertainty, but perform simulations at 1-km horizontal grid spacing, which allows us to perform many more experiments than would be possible using higher resolution.

Convection is initiated by inserting a cold pool in part of the domain at the start of the integrations, similar to the approach of Weisman et al. (1997). The initial cold pool perturbation potential temperature $\theta'$ linearly decreases with height, from −5 K at 0 km to 0 K at 4.5 km, and is inserted at $0 \leq x \leq 300$ km (where $x$ is the distance along the $x$ direction between 0 and 612 km). The initial cold pool extends across the domain in the $y$ direction. The magnitude of the lowest-level initial $\theta'$ is similar to or somewhat smaller than that of the cold pool that subsequently develops with the squall line. Random $\theta$ perturbations with maximum amplitude of ±0.05 K are applied between 275 < $x$ < 300 km and within the lowest 4.5 km to initiate motion in the $y$ direction.

The environmental sounding for all simulations is based on observations from Norman, Oklahoma, at 0000 UTC 20 June 2007 (Fig. 2), about 6 h prior to passage of the storm at this location. The initial $\theta$ and water vapor mixing ratio $q_v$ from the sounding are vertically smoothed using a 1.25-km running average (gray lines in Fig. 2). For simplicity, we idealize the vertical wind shear profile to give moderate unidirectional shear (in the $x$ direction) of 0.0048 s$^{-1}$ between 0 and 2.5 km, corresponding to a 12 m s$^{-1}$ difference in horizontal wind between these levels, with zero wind above. This is somewhat stronger than the observed low-level shear from the 0000 UTC sounding, which is nearly unidirectional at low levels, but with a difference in wind speed of only about 6–8 m s$^{-1}$ between the near-surface and 2.5 km (Fig. 2). Simulations with weaker low-level environmental shear closer to that observed by the sounding are unable to sustain a squall line (see section 4e). While this is perhaps suggestive of the model’s inability to simulate a squall line under realistic conditions, it is noted that the shear at the time of storm passage (~6 h prior to its passage) may have been stronger than observed at 000 UTC. Wind speed of the environment in the $y$ direction is set to zero. Sensitivity to the low-level vertical shear is tested, given its uncertainty and importance in storm structure and evolution. All of the breakup tests described...
above were run with either low (0.0032 s\(^{-1}\)), baseline, or high (0.0064 s\(^{-1}\)) vertical shear between 0 and 2.5 km (see Table 1). These results are detailed in section 4e.

Given initiation of the storm using an idealized linear cold pool, we focus mostly on comparing the simulated and observed storms several hours after their formation, when they have a similar width (perpendicular to the line) and exhibit clear convective, transition, and trailing stratiform regions. This allows for a more robust analysis that is less influenced by initial conditions. However, we note that comparison of the idealized modeled storm with specific times and locations of the observed storm is inevitably somewhat subjective, and this should be kept in mind when interpreting results.

4. Results

a. Comparison of baseline simulation and observations

Evolution of the reflectivity structure is shown for the BASE at a height of 1.13 km (Fig. 4). Vertical cross sections of reflectivity averaged in the \(y\) direction along the leading edge of the storm (defined by the lowest level \(-1\ K\ \theta^e\) isotherm) are shown in Fig. 5. Convection is initiated along the initial cold pool edge within the first 20 min. A region of stratiform precipitation develops after about 3–4 h, and the storm reaches a quasi-steady mature state after roughly 6 h, although the region of stratiform precipitation continues to widen over time.

The model reproduces the overall observed reflectivity structure (Fig. 6), with a well-defined convective line, transition region, and trailing stratiform region characteristic of midlatitude squall lines (e.g., Biggerstaff and Houze 1991). The simulated line-averaged reflectivity vertical structure also has similar features as observed; however, the observed 35-dBZ contour of the convective line reached 8 km at the mature stage of the MCS, compared to only 6 km in the model, and the convective line had reflectivity greater than 45 dBZ up to 6 km in the observed convective line, but only up to 4 km in the model. The observed bright band was not very well defined, and was centered at 3.5 km, while it has higher reflectivity in the model and is centered around 4 km.
These differences may be due in part to large uncertainty in the calculation of brightband reflectivity from model hydrometeor fields. The simulated reflectivity above (below) the bright band is also higher (lower) than observed in the stratiform region. The across-line extent of the squall line was observed to be about 190 km at this stage in its evolution, which is similar to the modeled extent of 170 km for BASE at 8 h.

Comparison of BASE with radar-retrieved and disdrometer-derived raindrop $D_0$ indicates that the model produces values that are consistently too large within the stratiform region by about a factor of 2 near the surface (Fig. 7), despite the low bias in reflectivity. The model also produces a sharp decrease of $D_0$ with height that is not seen in the radar retrievals. This is consistent with excessive drop size sorting during sedimentation, which is a known problem in two-moment schemes that assume exponential DSDs (Wacker and Seifert 2001; Milbrandt and McTaggart-Cowan 2010; Mansell 2010). This issue is further explored with a sensitivity test in which the number-weighted raindrop fall speed, which is applied to the sedimentation of rain number mixing ratio, is set equal to the mass-weighted fall speed applied to sedimentation of mass mixing ratio. This test prevents size sorting.
from occurring altogether, and produces a nearly constant $D_0$ with height in the stratiform region (not shown). These results therefore provide evidence that excessive size sorting may indeed be responsible for the unrealistic vertical profiles of $D_0$, although without size sorting the model still produces values of $D_0$ that are too large. We note that allowing $\mu$ to vary from an exponential distribution ($\mu = 0$) can improve the treatment of size sorting in two-moment schemes (Milbrandt and McTaggart-Cowan 2010). However, disdrometer observations indicate that the DSDs did not deviate greatly from exponential in the stratiform region for this storm, with fitted values of $\mu$ having considerable variability but generally ranging between 0 and 2 (with a tendency for larger $\mu$ with lighter precipitation in the transition region). That said, DSDs above the surface were not measured and may have deviated more substantially from exponential. Further investigation of this issue is beyond the scope of this paper and left for future work.

![Vertical cross section of line-averaged raindrop median volume diameter $D_0$ from (a) the BASE simulation at 8 h and (b) radar retrieval at 0750 UTC. (c) Comparison of disdrometer-derived $D_0$ and line-averaged $D_0$ from various simulations at the lowest model level. OBS1 and OBS2 indicate $D_0$ from disdrometer assuming exponential and gamma DSDs, respectively.](image)

**Fig. 7.** Vertical cross section of line-averaged raindrop median volume diameter $D_0$ from (a) the BASE simulation at 8 h and (b) radar retrieval at 0750 UTC. (c) Comparison of disdrometer-derived $D_0$ and line-averaged $D_0$ from various simulations at the lowest model level. OBS1 and OBS2 indicate $D_0$ from disdrometer assuming exponential and gamma DSDs, respectively.

![Comparison of (a) line-averaged and (b) probability density function (PDF) of modeled and radar-retrieved precipitation rates. Modeled precipitation rates are between 7 and 8 h and interpolated to a height of 1.13 km, and radar values are from 0750 UTC at the same height.](image)

**Fig. 8.** Comparison of (a) line-averaged and (b) probability density function (PDF) of modeled and radar-retrieved precipitation rates. Modeled precipitation rates are between 7 and 8 h and interpolated to a height of 1.13 km, and radar values are from 0750 UTC at the same height.

Modeled values of $D_0$ that are too large imply that the low bias of stratiform reflectivity below the melting level is primarily a result of too little rainwater as opposed to mean drop sizes that are too small. This is further suggested by large underprediction of modeled precipitation rates (at a height of 1.13 km) compared to radar retrievals (Fig. 8a). BASE produces a peak
precipitation rate in the convective region about 40% as large as that from radar, while precipitation rates in the stratiform region are 4–8 times smaller. A probability density function (PDF) of precipitation rate (Fig. 8b) shows modeled values to be nearly exponential, while radar rates show significant deviations from exponential, with relatively greater frequency of occurrence near 3–5 and 40–80 mm h⁻¹ and lower frequency at moderate rates. Here, the PDF is normalized to the total number of points with rain rates greater than 1 mm h⁻¹. These two regions of relatively higher frequency in the observed PDF correspond to the large region of stratiform rain and sharply peaked region of convective rain with high precipitation rates; these regions are less distinct in the simulations.

b. Sensitivity to parameterization of raindrop breakup

Additional simulations were performed to test sensitivity of the squall line to parameterization of raindrop breakup.
breakup, as described in section 3b and summarized in Table 1. First, we describe tests using the S08 and Z85 breakup parameterizations. S08 and Z85 also treat breakup by reducing rain self-collection efficiency $E_c$ similar to the modified VC93 formulation in BASE expressed by (4), but these schemes produce large differences in $E_c$ (see Fig. 3). Different breakup formulations produce widely varying rain DSD parameters (Fig. 9), corresponding with large differences in their impact on $E_c$. Of the three schemes that were tested (BASE, S08, and Z85), Z85 produces the least efficient drop breakup overall and therefore the smallest values of $N_0$ for a given rainwater content RWC especially for RWC > 0.1 g m$^{-3}$. BASE and S08 produce similar mean relationships between $N_0$ and RWC, although BASE produces a somewhat greater spread for a given RWC. Thus, overall breakup efficiency appears to be similar between BASE and S08, with less efficient breakup (larger $E_c$) in BASE compared to S08 at smaller $D_r$ compensated by more efficient breakup in BASE for $D_r > 0.6$ mm (see Fig. 3). Simulations using any of the three breakup formulations underpredict $N_0$ by one to two orders of magnitude for a given RWC compared to disdrometer. Since $D_0$ is proportional to $N_0^{1/4}$, this bias of $N_0$ corresponds to values of $D_0$ that are too large relative to disdrometer and radar retrievals. There is a close correspondence between the $N_0$–RWC relationship for this case and the 2007 seasonal observations collected in deep convective systems (comprising 15 cases; Fig. 10). Thus, it appears that the 20 June case provides a representative sample of DSD characteristics, and the model biases cannot be explained by anomalous characteristics for this particular case.

Time series of various quantities are shown in Fig. 11 for BASE, S08, and Z85. This analysis includes domain-mean surface precipitation rate, total condensation rate, total evaporation rate, precipitation efficiency (defined as total condensation rate divided by surface precipitation rate), average lowest-level $\theta'$ within the cold pool, fraction of the domain at the lowest level with cold pool, ratio of cold pool intensity ($C$) to the difference in the 0 and 2.5 km horizontal wind ($\Delta U$), and mean propagation speed of the surface gust. (Hereafter, “total” refers to vertically integrated and domain-averaged quantities.) Throughout this paper, the cold pool is defined by the region with $\theta' \leq -2$ K.

Here, $C$ is defined using (e.g., Weisman and Rotunno 2004; Bryan et al. 2006)

$$ C^2 = 2 \int_0^H (-B) \, dz , \quad (5) $$

where $H$ is the cold pool depth (defined as the height of the level of neutral buoyancy), and $B$ is buoyancy.

$$ B = g \left[ \frac{\theta - \overline{\theta}}{\theta} + 0.61(q_v - q_h) - q_h \right] . \quad (6) $$

Here $g$ is acceleration of gravity, $q_h$ is the hydrometeor mass mixing ratio, and overbars denote the model’s base state. For our analysis, $C$ is averaged over the region 30 km behind the surface gust front.

Overall, the three breakup parameterizations produce fairly similar results. Domain-mean surface precipitation during the mature phase of the storm (from 6 to 8 h) differs by about 10% among the simulations, although earlier differences exceed 100% (between ~3 and 5 h). These differences can be partially explained by precipitation efficiency, which is highest in Z85 owing to the larger mean drop size and hence fall speed, and lowest in BASE. There are no consistent differences in total condensation, evaporation, cold pool area and strength, propagation speed, or updraft characteristics between the simulations.

These results broadly suggest limited sensitivity of the storm to drop breakup for the three schemes that were tested. However, there is much greater sensitivity using the modified VC93 scheme as in BASE, but varying the size threshold for breakup $D_{th}$. Here, $D_{th}$ is varied between 105 and 510 $\mu$m, compared to 300 $\mu$m for BASE (see Table 1). Setting $D_{th}$ to 105 $\mu$m (BREAK1) produces values of $N_0$ that are similar to observations for RWC greater than about 0.1 g m$^{-3}$ (Fig. 9). BREAK1 also gives more realistic values of $D_0$ than BASE when...
compared to disdrometer, although values are still somewhat too large in the trailing stratiform region (Fig. 7c). Increasing $D_{th}$ to 405 μm (BREAK2) and 510 μm (BREAK3) results in progressively smaller $N_0$ (larger $D_0$) for a given RWC, and increases the bias of $N_0$ (and $D_0$) relative to observations (Figs. 7c and 9).

Time series of various quantities (as in Fig. 11) are shown in Fig. 12 for BREAK1, BASE, BREAK2, and BREAK3. These simulations have considerably more spread than BASE, S08, and Z85, especially before 7 h. In general, there is a tendency for greater domain-mean surface precipitation with more efficient drop breakup (Fig. 12a; in contrast with peak surface precipitation rates as described below). This is unexpected, since more efficient breakup leads to higher evaporation rates and lower mass-weighted fall speeds for a given RWC. Overall, there is a close correspondence between rain evaporation and surface precipitation in these simulations (Figs. 12a,c), with larger domain-mean precipitation rates occurring with larger total evaporation rates. For example, BREAK1 has by far the largest domain-mean precipitation rates (especially before 6 h), despite the fact that it

Fig. 11. Time series of (a) domain-mean surface precipitation rate, (b) total condensation rate, (c) total evaporation rate, (d) domain-mean precipitation efficiency, (e) mean lowest-level $\theta^\prime$ within the cold pool, (f) cold pool area, (g) ratio cold pool intensity and difference in 2.5- and 0-km wind $C/\Delta U$, and (h) propagation speed for the BASE, S08, and Z85 simulations.
produces relatively small drops and hence lower mass-weighted fall speeds and higher evaporation for a given RWC. We propose the following chain of interactions to explain this counterintuitive result. First, larger evaporation rates lead to greater cold pool intensity (Figs. 12e–g) and hence faster propagation of the surface gust front (Fig. 12h), yielding a wider storm over time. A larger storm results in greater domain-mean updraft mass flux and hence total condensation (Fig. 12b), which compensates for the higher evaporation rates and leads to the increased domain-mean surface precipitation. Furthermore, greater condensation implies more condensed water is available for evaporation, leading to even stronger cold pools, faster propagation, etc. This represents a positive feedback that can enhance the sensitivity of modeled convective storm features to microphysical processes. van Weverberg et al. (2012) also found increases in total condensation (latent heating) with stronger cold pools in two-dimensional squall-line simulations, although they reported less sensitivity of surface precipitation because of compensating changes in precipitation efficiency.

Horizontal cross sections of reflectivity at 8 h and a height of 1.13 km for BREAK1, BREAK2, and BREAK3 simulations (Fig. 13) reflect storm propagation that is about a factor of 2 slower in BREAK2 than BREAK1 (Fig. 12h), leading to a relatively narrow storm. At this time, the storm simulated in BREAK1 is approximately twice as wide as that in
BREAK2. Between 5 and 7 h, the storm in BREAK2 loses its linear organization and becomes highly asymmetric in the y direction, before reorganizing as a contiguous line after $\sim$7 h. While width and organization are the most apparent differences in terms of overall structure, other notable differences include relatively weak low-level reflectivity in the stratiform region in BREAK1 and higher reflectivity in BREAK2 and BREAK3. While more efficient breakup in BREAK1 improves comparison of the DSD parameters ($N_0$ and $D_0$) with disdrometer and radar compared to the other simulations, it leads to an even greater low bias of low-level reflectivity in the stratiform region compared to radar (cf. Figs. 6a,c,e and 13).
Plots of line-average precipitation at a height of 1.13 km indicate sensitivity of peak precipitation rates to drop breakup (Fig. 8a), consistent with van Weverberg et al. (2012). In contrast with domain-mean precipitation, peak precipitation rates increase with a reduction in drop breakup efficiency as a result of the increase in mean drop size and hence mass-weighted fall speed. PDFs of precipitation rate show relatively more instances of low-precipitation rates in BREAK1, but fewer at middle and high rates (greater than about 10 mm h$^{-1}$), especially compared to BREAK2 (Fig. 8b).

BREAK2 has a much smaller domain-mean updraft mass flux than the other simulations (Fig. 14a), which is consistent with its relatively small storm area (Fig. 13). However, it has stronger convective updrafts above ~4.5 km (Fig. 14b). Here convective updrafts are defined using a threshold vertical velocity $w$ of 2 m s$^{-1}$, although overall results are not sensitive to this threshold. The fraction of convective updrafts is also relatively small in BREAK2 below 8 km, but larger above this height (Fig. 14c), implying that updrafts tend to reach higher levels in this simulation. This presents an overall picture of fewer and/or smaller but stronger and more upright updrafts in BREAK2 compared to the other runs. Conversely, BREAK1 has the greatest domain-mean updraft mass flux below 5 km, but a relatively small fraction of convective updrafts above 8 km. These differences are generally consistent with peak vertical velocities over the domain (not shown). BREAK2 and BREAK3 tend to produce the strongest peak updrafts, while BREAK1 tends to have the weakest, although there is considerable variability. PDFs of updraft velocity at a height of ~7.5 km (Fig. 15) also show that BREAK2 tends to have relatively more strong updrafts ($w > 10$–$15$ m s$^{-1}$) and fewer weak updrafts ($w < 5$ m s$^{-1}$), with the opposite for BREAK1.

Interestingly, the modeled storm exhibits a clear nonmonotonic sensitivity to $D_{th}$. For BREAK1, BASE, and BREAK2, the increase in $D_{th}$ generally results in less domain-mean surface precipitation, smaller and weaker cold pools, slower propagation, and narrower storms as described above. However, a further increase in $D_{th}$ from BREAK2 to BREAK3 has the opposite effects; results from BREAK3 are remarkably similar to BASE (as well as S08 and Z85). Increasing $D_{th}$ beyond the value in BREAK3 produces relatively limited change in results compared to BREAK3 (not shown). Thus, BREAK2 produces the lowest surface precipitation rate and weakest cold pool of any simulation, despite the fact that it lies in the middle range of breakup efficiencies that were tested. To explore robustness of this result, we performed additional simulations with small changes to $D_{th}$ in the vicinity of the value set for BREAK2. These simulations, with $D_{th}$ set to either 380 or 430 $\mu$m, are similar to BREAK2 (not shown). Thus, the nonmonotonic response appears to be robust.
This nonmonotonic response of the storm to drop breakup also likely explains greater sensitivity to varying $D_{th}$ than using the different breakup formulations (BASE, S08, and Z85). In terms of overall drop breakup efficiency, BREAK2 lies between BASE, S08, and Z85, as indicated by the $N_0$–RWC relationships in Fig. 9, yet it produces a much lower domain-mean surface precipitation rate, weaker cold pool, and slower propagation than any of these simulations. On the other hand, BREAK3 and Z85 have similarly inefficient drop breakup overall (as again suggested by the $N_0$–RWC relationships), yet both simulations produce similar storm characteristics compared to BASE and S08. Thus, BREAK2 happens to lie in a region of parameter space that produces distinctly different storm characteristics from BASE, S08, Z85, and BREAK3. Additional sensitivity tests are described in section 4c that show this nonmonotonic behavior is related to the impact of drop breakup on mean rain fall speed.

c. Impact of changes in raindrop fall speed

While the results in section 4b suggest that differences in evaporation and hence cold pool evolution and storm dynamics are critical in explaining sensitivity of the simulated storm to drop breakup, the nonmonotonic response to drop breakup suggests that additional factors are also important. In other words, if evaporation was the only controlling process, progressively more efficient drop breakup and hence smaller mean drop size and greater evaporation (for a given RWC) would lead one to expect a monotonic sensitivity to breakup. Hence, a key question is what explains the nonmonotonic response of the modeled storm to parameterization of drop breakup?

Additional sensitivity tests with modifications to the rain fall speed help to address this question. In the model, the mass- and number-weighted fall speeds are given by

$$V_m = \frac{a\Gamma(b + 4)}{6\lambda^b},$$  \hspace{1cm} (7)  

$$V_N = \frac{a\Gamma(b + 1)}{\lambda^b},$$  \hspace{1cm} (8)

where $a$ and $b$ are parameters in the power-law fall speed–size relation (see Morrison 2012). In these tests, $\lambda$ is formulated to be independent of the predicted mean drop size for calculation of $V_m$ and $V_N$, and therefore is not impacted directly by drop breakup (for all other processes, the model-predicted values of $\lambda$ are used). Thus, instead of using (2) to calculate $\lambda$ for (7) and (8), it is diagnosed as a function of RWC following:

$$\lambda = 2.6 \times 10^3 RWC^{-0.015},$$  \hspace{1cm} (9)

where $\lambda$ and RWC have units of m$^{-1}$ and g m$^{-3}$, respectively. The $\lambda$–RWC relationship in (9) produces values of $\lambda$ between BREAK1 and the disdrometer observations for a given RWC; however, the specific formulation of $\lambda$ from (9) is less important than the fact that the same $\lambda$–RWC relationship is used in all of these sensitivity tests for calculating $V_m$ and $V_N$. As in section 4b, these simulations are performed with $D_{th} = 105, 300, 405, \text{ and } 510 \mu$m (BREAK1F, BASEF, BREAK2F, and BREAK3F, respectively; see Table 1), along with baseline (moderate) low-level environmental wind shear.

With $V_m$ and $V_N$ independent of the predicted mean drop size and hence breakup, quantities such as surface precipitation rate, total evaporation and condensation, cold pool intensity, propagation speed (Fig. 16), and storm width and structure (not shown) exhibit a generally monotonic response to $D_{th}$, especially before 6 h, in contrast with results described in section 4b. In these runs, progressively less efficient drop breakup generally leads to reduced evaporation, weaker cold pools, slower propagation, narrower storms, reduced condensation, and reduced surface precipitation, with no evidence for a strongly nonmonotonic response to $D_{th}$, especially before 6 h, in contrast with results described in section 4b. In these runs, progressively less efficient drop breakup generally leads to reduced evaporation, weaker cold pools, slower propagation, narrower storms, reduced condensation, and reduced surface precipitation, with no evidence for a strongly nonmonotonic response to $D_{th}$, especially before 6 h, in contrast with results described in section 4b. These simulations are performed with $D_{th} = 105, 300, 405, \text{ and } 510 \mu$m (BREAK1F, BASEF, BREAK2F, and BREAK3F, respectively; see Table 1), along with baseline (moderate) low-level environmental wind shear.

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speed on the response to breakup are unclear but appear to be related to complex interactions between sedimentation and loading of rainwater, evaporation, cold pool dynamics, and environmental wind shear. The importance of environmental wind shear is further supported by tests described in section 4c; simulations with increased low-level shear also produce a monotonic sensitivity to breakup. Further work is needed to investigate these interactions systematically.

d. Implications for RKW theory

Close correspondence between the convective strength and cold pool intensity in these simulations broadly support “RKW theory” (Rotunno et al. 1988; Weisman and Rotunno 2004). According to this theory, an optimal state for system intensity occurs when the vorticity associated with low-level environmental shear is approximately balanced by the cold pool’s circulation (i.e., when $C/\Delta U \sim 1$), leading to the deepest lifting of environmental air (all else being equal) and upright convective cells. All simulations here give $C/\Delta U > 1$ (Fig. 12g), leading to an upshear-tilted convective structure. However, BREAK1, with the highest evaporation rates, produces the strongest cold pool and hence largest $C/\Delta U$, especially before 6 h. Consistent with RKW theory, there are fewer strong updrafts ($w > 10-15$ m s$^{-1}$),
weaker convective mass flux above ~5 km, and shallower, more tilted convective updrafts (i.e., relatively low convective updraft fraction above 8 km) in this simulation. The opposite occurs in simulations with less evaporation and relatively smaller $C/\Delta U$ (closer to unity), in particular BREAK2.

While the simulations in a less optimal state (i.e., $C/\Delta U$ farther from unity) with high evaporation rates produce relatively weaker convective updrafts, other measures show an increase in storm intensity. In particular, storm size, domain-mean updraft mass flux, total condensation rate, and mean surface precipitation are all greater in the simulations with higher evaporation and larger $C/\Delta U$, because of the interactions between microphysics, cold pool evolution, storm size, and total condensation discussed in section 4b. In a multimodel assessment of RKW theory applied to squall lines, Bryan et al. (2006) also reported greater total condensation in simulations with relatively less optimal conditions. However, in contrast to our results, they found that total surface precipitation was lower in these simulations.

e. Sensitivity to vertical wind shear

Given the importance of interactions between microphysics, cold pool evolution, propagation speed, and storm dynamics, low-level vertical wind shear is expected to play an important role in the response of the storm to drop breakup. This idea is broadly consistent with the importance of shear on squall-line intensity and structure in previous squall-line simulations (e.g., Weisman and Rotunno 2004) and encapsulated by RKW theory. To explore this issue, simulations with varying $D_{th}$, as described in previous subsections for baseline (moderate) shear, were performed with the $0$-2.5-km shear either decreased from $0.048 (\Delta U = 12 \text{ m s}^{-1})$ to $0.0032 \text{ s}^{-1} (\Delta U = 8 \text{ m s}^{-1})$ or increased to $0.0064 \text{ s}^{-1} (\Delta U = 16 \text{ m s}^{-1})$. Regardless of the breakup parameterization, all simulations with the weaker shear failed to produce a long-lived squall line, with initial convection dissipating within the first few hours and only weak, isolated convective cells remaining.

Simulations with higher shear (BREAK1H, BASEH, BREAK2H, and BREAK3H; see Table 1) produce long-lived squall lines (Fig. 17), but give results that are substantially different than with baseline shear (cf. Figs. 17 and 12). In particular, the higher shear simulations produce more precipitation, especially before 6 h, and stronger cold pools than those with baseline shear. Despite the stronger cold pools, the increase in cold pool intensity $C$ is not enough to compensate for the larger $\Delta U$, and hence $C/\Delta U$ is smaller (i.e., closer to unity) at high shear compared to baseline shear, mainly after 4–5 h. Overall, there is greater sensitivity to shear than drop breakup for the parameter range tested here. Large sensitivity of the organization and structure of squall lines to shear is expected and has been well documented in previous modeling studies (e.g., Thorpe et al. 1982; Nicholls et al. 1988; Weisman et al. 1988; Fovell and Ogura 1989; Szeto and Cho 1994; Robe and Emanuel 2001; Weisman and Rotunno 2004; James et al. 2005; Bryan et al. 2006; Takemi 2007). Despite large sensitivity of storm structure, there is limited impact of shear on the rain DSD characteristics, such as the $N_{0}-$RWC relationship (not shown).

Higher shear simulations exhibit similar relationships between total evaporation, cold pool intensity, propagation speed, total condensation, and domain-mean surface precipitation as in the baseline shear simulations, although overall sensitivity to breakup is smaller. These results again highlight the dominant control of interactions between microphysics, cold pool evolution, and dynamics on the response of domain-mean surface precipitation to drop breakup, rather than the direct impact of raindrop size on precipitation. Thus, despite its tendency to produce the smallest mean drop size, and hence lowest fall speed and highest evaporation rate for a given RWC, BREAK1H produces the highest domain-mean precipitation rate for the high shear runs (Fig. 17a). This is consistent with the interactions between microphysics, cold pool evolution, and storm dynamics described in section 4b.

An important difference from the baseline shear simulations is that the response to drop breakup under higher shear is monotonic; that is, various storm characteristics tend to have a monotonic change as $D_{th}$ is increased (Fig. 17). This qualitative difference in the response, depending on shear, further supports our suggestion that the nonmonotonic response under baseline shear depends on interactions between drop sedimentation, evaporation, cold pool dynamics, and environmental shear.

5. Summary and conclusions

This paper describes a series of idealized 3D simulations of a squall line observed on 20 June 2007 over central Oklahoma. The baseline model configuration reproduced key qualitative storm features, including a leading edge of high reflectivity, a wide region of trailing stratiform precipitation, and a transition region of low reflectivity between the stratiform and convective regions. However, quantitatively the model exhibited a low bias of reflectivity in the stratiform region even though raindrop median volume diameter was too large compared to radar retrievals and disdrometer measurements, indicating too little rainwater. It also underpredicted precipitation rates in the convective region by a factor of roughly 2.5, and
in the stratiform region by a factor of 4–8. Reasons for the large underprediction of precipitation are unclear, but could be due to uncertainties in the environmental conditions (e.g., sounding) or model configuration (microphysics or other aspects). Further work based on longer-term statistical comparisons with observations is needed to determine if this bias is a consistent feature of the microphysics scheme.

Different formulations for drop breakup from the literature (BASE, S08, and Z85), as well as varying threshold mean drop size $D_{th}$ for breakup using the modified VC93 formulation, were tested under different 0–2.5 km vertical wind shears. Rain DSD characteristics, such as the $N_0$–RWC relationship, were insensitive to environmental shear. Simulations with BASE, S08, and Z85 produced values of rain $N_0$ that were too small by one–two orders of magnitude for a given RWC, with a corresponding large bias in mean drop size. Thus, these formulations appear to produce breakup efficiencies that are too weak, although we cannot rule out DSD biases resulting from other processes such as excessive size sorting, which is a known problem in two-moment schemes with exponential DSDs (Wacker and Seifert 2001; Milbrandt and McTaggart-Cowan 2010; Mansell 2010). A sensitivity test with mass-weighted raindrop fall speed applied to sedimentation of both the rain number and mass mixing

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**FIG. 17.** As in Fig. 11, but for the BREAK1H, BASEH, BREAK2H, and BREAK3H simulations with higher vertical wind shear.
ratios, thereby preventing size sorting from occurring altogether, suggested that excessive size sorting may have indeed been responsible for an unrealistic decrease of $D_0$ with height in the stratiform region in the baseline simulation. However, without size sorting the model still produced mean drop sizes that were too large compared to radar retrievals. Decreasing $D_{th}$ from 300 $\mu$m in BASE to 105 $\mu$m in BREAK1 improved comparison of the DSD parameters with observations. Gao et al. (2011) also found an overprediction of mean drop size ($Z_{DR}$) using their two-moment scheme, and improvement of DSD characteristics by increasing the breakup efficiency. However, more efficient breakup in BREAK1 tended to degrade other aspects such as maximum (line averaged) surface precipitation and reflectivity structure. These results suggest that the primary model biases of low surface precipitation and low-level reflectivity within the stratiform region were not a result of misrepresentation of the rain DSDs and hence breakup.

We note that other aspects of the treatment of rain DSDs may also be important. For example, the parameterization of DSD shape or width (i.e., $\mu$ for the gamma DSD) impacts processes such as rain evaporation and sedimentation, and size sorting in particular (Milbrandt and McTaggart-Cowan 2010), but we leave this as a subject for future work. Disdrometer observations can also provide useful constraints on the parameterization of $\mu$ (e.g., Cao et al. 2008), and we plan to pursue such efforts in a follow-up study. We note, however, that existing observationally based parameterizations of $\mu$ have relied upon surface disdrometer measurements; observational characterization of rain DSDs above the surface are needed to test the generality of these parameterizations.

A key finding of this study is that the response of domain-mean surface precipitation to drop breakup by varying $D_{th}$ was mostly controlled by interactions between the microphysics, cold pool characteristics, and dynamics, rather than direct impacts of breakup on drop size and precipitation as one might have expected. Thus, simulations with more efficient breakup tended to have higher domain-mean precipitation rates (but lower peak precipitation rates), despite the smaller mean drop size and hence lower fall speed and higher evaporation rate for a given RWC, under both baseline moderate (0.0048 s$^{-1}$) and high (0.0064 s$^{-1}$) low-level environmental wind shears. We propose that this countervuitive result occurred because higher evaporation led to stronger cold pools, faster propagation, larger storm area, greater domain-mean updraft mass flux, and greater total condensation that compensated for higher evaporation to give more, not less, precipitation. This picture is complicated by the impact of drop breakup on mass- and number-weighted fall speeds, which was found to be critical in explaining the nonmonotonic response of storm characteristics to drop breakup under baseline wind shear. For higher wind shear, the response was monotonic; more efficient breakup generally led to progressively greater evaporation, stronger cold pools, faster propagation, and more surface precipitation. Under weaker shear (0.0032 s$^{-1}$), a squall line could not be maintained using any of the breakup schemes. Our simulations used only a single thermodynamic profile from a midlatitude, continental environment. Future work should evaluate the generality of these findings for other conditions, such as tropical maritime environments with lower CAPE and significantly higher low-level relative humidities.

Overall, interactions between microphysics (evaporation in particular), cold pool intensity, and low-level environmental wind shear in the simulations broadly support RKW theory (Rotunno et al. 1988; Weisman et al. 1988). Simulations with relatively strong cold pools and $C/\Delta U$ farthest from unity (and hence farthest from “optimal” vorticity balance, which occurs when $C/\Delta U \approx 1$) had weaker, more tilted convective updrafts at mid and upper-levels that tended not to penetrate as high as the more intense, upright updrafts in simulations with weaker cold pools, consistent with RKW theory. However, other measures of storm strength, including domain-mean surface precipitation, overall storm size and hence mean updraft mass flux, and total condensation, were significantly greater in the simulations with stronger cold pools and larger $C/\Delta U$, especially earlier in the simulations. Bryan et al. (2006) also found greater total condensation in less optimal conditions in their multimodel study, but surface precipitation was larger in simulations with a more optimal balance. A possible explanation for the contrasting results is our use of a relatively sophisticated microphysics scheme that includes both liquid and ice hydrometeors; most previous numerical evaluations of RKW theory (e.g., Weisman et al. 1988; Weisman and Rotunno 2004; Bryan et al. 2006) used liquid-only schemes (e.g., Kessler 1969), which are known to have difficulty simulating realistic stratiform precipitation (e.g., Fovell and Ogura 1988). Our simulations produced large stratiform regions that contributed significantly to the total condensation, evaporation, and cold pool evolution, and appeared to be critical in driving the aforementioned interactions between microphysics, cold pool evolution, and dynamics that largely control the response of the domain-mean surface precipitation rate to changes in microphysics. We also note that Bryan et al. (2006) assessed RKW theory by modifying low-level environmental shear but not cold pool intensity, while we
performed tests with changes to drop breakup that impacted cold pool intensity. Further work is needed to investigate more systematically the microphysical context of RKW theory in squall lines across a greater range of shears.

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**APPENDIX**

**Description of the Observations**

*a. Disdrometer observations*

Detailed descriptions of the two-dimensional video disdrometer (2DVD) and processing techniques can be found in Kruger and Krajewski (2002), Brandes et al. (2007), and Cao et al. (2008). The 2DVD has two high-resolution cameras orthogonal to one another and records silhouette images of raindrops falling through a 10 cm × 10 cm measurement area from which equivalent volume diameter, oblateness, and terminal velocity of each raindrop are derived. In this study, the disdrometer data were quality controlled by ignoring any particles whose measured terminal velocity significantly deviated from the wind tunnel raindrop terminal velocity. The data were then sorted into 41 size bins of 0.2-mm width and having central diameters of 0.1–8.1 mm to determine 1-min DSDs. The parameters of the gamma distribution were computed from the second, fourth, and sixth moments of the drop distributions following the moment-fitting approach of Ulbrich and Atlas (1998). For the exponential distribution, the DSD parameters (such as $N_0$ in Figs. 9 and 10) were calculated from the third and sixth moments. The disdrometer-derived raindrop median volume diameter $D_0$ (Fig. 7c) was calculated directly from either the fitted gamma or exponential DSDs.

*b. Radar analysis procedures*

The KOUN radar data were cleaned up and gridded prior to the analysis. Ground clutter and nonmeteorological data were omitted by removing data that had a correlation coefficient ($\rho_{hv}$) less than 0.85. The correlation coefficient is a polarimetric radar variable that indicates the level of correlation between the horizontal and vertical polarized backscattered power in a sample volume, and can be a good indicator of nonmeteorological targets when there is low correlation in the sample volume (i.e., hydrometeor classification categories are typically associated with $\rho_{hv} > 0.85$; Straka et al. 2000; Schuur et al. 2003; Liu and Chandrasekar 2000; Park et al. 2009). This procedure also removed spurious data due to second trip echo contamination. Second trip echoes (radar echoes received from a previous pulse volume after the next pulse has been transmitted) were triggered by the extensive and intense squall line that extended to the southwest (and outside) of the KOUN radar domain. Radar-derived variables (rain rate and $D_0$, see discussion below) were calculated from the radar data in polar coordinates after the data were cleaned up. Finally, the equivalent radar reflectivity ($Z_{HH}$), retrieved rain rate, and $D_0$ were gridded to a Cartesian coordinate system with 0.5-km spacing using linear interpolation.

To compare with the model simulation results, the gridded data were rotated by 55° to place the majority of the convective line perpendicular to the $x$ axis. Line-averaged calculations of $Z_{HH}$, $D_0$, and retrieved rain rate (taken at 1.13 km AGL) were then performed to represent an average cross section of the squall line. This was done using a 50-km-wide slab of data 25–75 km east of the radar (on the rotated grid) to avoid influence of the cone of silence. Choosing other locations to define the slab for cross-section averaging (e.g., west of the radar) was tested and did not impact overall results or conclusions.

c. Methodology for radar rain rate and $D_0$ retrievals

Numerous studies have shown that dual-polarization radar methods to compute rain rates perform better compared to calculating rain rates from $Z_{HH}$ alone, because dual-polarization radar measurements are sensitive to particle size, shape, orientation, phase, and bulk density (e.g., Goddard and Cherry 1984; Ryzhkov and Zrnić 1995; Brandes et al. 2002; Brandes et al. 2003). In this study, following the technique of Brandes et al. (2002) and Brandes et al. (2003), radar-based rain rates were computed from three combinations of dual-polarization radar variables: 1) $Z_{HH}$ and differential reflectivity ($Z_{DR}$), 2) specific differential phase ($K_{DP}$) and $Z_{DP}$, and 3) $K_{DP}$ alone. This technique starts by utilizing rain DSDs from
disdrometer observations to compute $Z_H$, $Z_{DR}$, and $K_{DP}$ as described by Zhang et al. (2001). Power-law rainfall estimators (one for each of the combinations of dual-polarization radar variables listed above) are then determined by the least squares method with the disdrometer measurements of rain rate as the dependent variable. The three power-law rainfall estimators used in this study come from more than 4300 one-minute DSDs collected by the 2DVDs in Oklahoma during stratiform and convective rain events during the summer seasons of 2005–07.

To generate a single rain-rate field for each radar scan based on the three rain-rate estimators, a decision-making algorithm (referred to as “hybrid” rain-rate algorithm hereafter) was used on a gate-by-gate basis. The hybrid algorithm uses the rain rates from the $Z_H$–$Z_{DR}$ combination or from $K_{DP}$ alone in moderate and heavy (convective) rain, respectively, while rain rates from the $K_{DP}$–$Z_{DR}$ combination are used for light (stratiform) rain. Compared with the rain rates observed at Oklahoma Mesonet stations within the radar sample domain between 0400 and 1100 UTC 20 June 2007, the hybrid rain-rate algorithm produced a small mean bias of 0.09 mm h$^{-1}$ (with a standard error of 0.98 mm h$^{-1}$). The mean bias was $-1.37$ mm h$^{-1}$ ($1.66$ mm h$^{-1}$) for the $Z_H$–$Z_{DR}$ rain rates, 0.55 mm h$^{-1}$ ($1.30$ mm h$^{-1}$) for the $Z_{DR}$–$K_{DP}$ rain rates, and 0.55 mm h$^{-1}$ ($1.30$ mm h$^{-1}$) for the $K_{DP}$ rain rates (values in parentheses indicate standard error).

The radar-retrieved $D_0$ are based on a relationship between $Z_{DR}$ and $D_0$ obtained from the 1-min disdrometer data collected between 2005 and 2007 and fitted to gamma DSDs. The derived $Z_{DR}$–$D_0$ relation was then applied to the $Z_{DR}$ field measured by KOUN to produce the radar-based $D_0$ field discussed in section 4 (Fig. 7b). To verify the radar-based $D_0$ for the 20 June 2007 squall line, the values within 3 km of the National Center for Atmospheric Research (NCAR) disdrometer site were averaged and compared to $D_0$ from the disdrometer observations. This comparison indicates good agreement between the disdrometer and radar-derived $D_0$ for this case.

The KOUN radar transmits and receives horizontal and vertical polarized electromagnetic waves simultaneously (hereafter SHV), which has been shown to result in cross coupling of the orthogonal waves due to depolarization of canted ice crystals near the tops of thunderstorms (Ryzhkov and Zrnic 2007). This artifact was observed above the melting level in this particular case, and given there is no correction for this issue we derive rain rates and $D_0$ from polarimetric measurements only below the melting level ($\sim 3.5$ km). Moreover, recent studies (Hubbert et al. 2010a,b; Zrnic et al. 2010) have documented that a potential bias in $Z_{DR}$ in rain may exist even below the melting level from cross coupling in SHV measurements compared to alternating transmit systems, which may affect polarimetric-derived rainfall rates and $D_0$. However, given the good agreement between the hybrid radar and mesonet rain rates and between the radar-derived and disdrometer $D_0$, it does not appear that a $Z_{DR}$ bias negatively affected retrievals of these quantities in this study.

REFERENCES


