Sensitivity of Tornadogenesis in Numerical Supercell Simulations to Variations in Microphysical Parameters

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Abstract

Idealized numerical simulations of tornadic thunderstorms are performed using the Advanced Regional Prediction System (ARPS), in which convection is initiated within an environment defined using a sounding associated with the May 20, 1977 tornadic supercell that occurred near Del City, Oklahoma. Thirteen simulations are conducted at 1 km horizontal grid spacing, and seven of them are repeated at the 100 m horizontal resolution, using various values of the intercept parameters for rain, snow, and hail for the drop size distributions found in a single-moment ice microphysics scheme. Influence of these parameters as well as the hail density on storm dynamics and tornadogenesis is analyzed.

It is found that the cold pool intensity and low-level storm dynamics are most sensitive to rain and hail intercept parameters, with much less sensitivity to snow intercept parameter and hail density. A microphysical budget analysis reveals that evaporation of rain and melting of graupel play the largest roles in cold pool intensification. Drop size distributions that favor smaller hydrometeors, particularly smaller raindrops lead to storms with a more intense cold pool due to enhanced evaporative cooling over a larger geographic area, while drop size distributions favoring larger hydrometeors produce weaker cold pools. Two of the seven 100 m simulations produce sustained surface tornadic circulations, with maximum intensity of F2 and durations of 4 minutes in the one with control settings and 9 minutes in the other with a reduced value of rain intercept parameter. In cases with very strong cold pools, the storm updraft is tilted rearward behind the gust front due to the rapid forward surge of the surface gust front, which often precedes the updraft maximum by several kilometers. Cases with weaker cold pools feature strong, sustained, vertical updrafts that are positioned directly over the low-level gust front and circulation maxima. The latter positioning results in a more intense, vertical updraft, being aided
by the upward vertical pressure gradient force dynamically induced by the presence of mid-level mesocyclone, and provide a favorable condition for tornadogenesis through enhanced low-level vertical stretching.
1. Introduction

With the recent improvement in numerical weather prediction (NWP) models and data assimilation techniques, together with the rapid increase in computational power, explicit prediction of both organized convective systems and individual convective storms has become a reality (e.g., Xue et al. 2003; Hu and Xue 2006; Hu et al. 2006). Over the continental United States, particularly the Central Great Plains, available data from dense surface networks and other platforms can be assimilated into numerical models at frequent intervals to achieve accurate prediction of the timing and location of convective initiation (e.g., Liu and Xue 2006; Xue and Martin 2006) while Doppler radar data can be assimilated, using advanced techniques such as 4DVAR or ensemble Kalman filter methods, into storm-scale models to produce reasonably accurate analysis or state estimation of pre-existing storms (e.g., Dowell et al. 2004; Sun 2005; Hu et al. 2006). As data assimilation techniques improve, the errors or uncertainties in the prediction model start to become a major issue. In fact, model errors affect both data assimilation (e.g., Tong and Xue 2006) and prediction.

For short-range convective-scale data assimilation and prediction, the model microphysics scheme appears to be the largest source of uncertainty, and thus potential error. Most commonly used microphysics schemes employ the 'bulk' approach of parameterization, in which the particle or drop size distributions (DSD’s) are parameterized in functional forms. Often, significant uncertainties exist with the treatment of microphysical processes and microphysical parameters. Previous sensitivity studies (e.g., McCumber et al. 1991; Ferrier et al. 1995; Gilmore et al. 2004; van den Heever and Cotton 2004; Tong and Xue 2006) demonstrate that the structure and evolution of simulated convective systems are very sensitive to microphysical parameterizations. Variations in microphysical parameters, such as collection
efficiencies, DSD parameters and particle densities, have profound effects upon the characteristics of precipitation systems and their associated dynamical processes.

Through idealized simulations of a supercell storm, Gilmore et al. (2004) showed that variations in DSD-related microphysical parameters within their range of uncertainty can cause significant changes in hydrometeor type, precipitation accumulation, and precipitation intensity while van den Heever and Cotton (2004) show that similar variations can cause a high-precipitation (HP) supercell storm to change into a low-precipitation (LP) storm. In both studies, a horizontal resolution of 1 km is used by the simulations; therefore an explicit simulation of tornadogenesis within the supercell storms cannot be expected. In this study, we perform a set of idealized numerical simulations of supercell storms, using a horizontal resolution of 100 m, sufficient to explicitly simulate the genesis and life cycle of larger tornado(s) within the supercell. We vary the DSD-related parameters within an ice microphysics scheme employed by the simulation model to examine the sensitivity of tornadogenesis to these microphysical parameters. Through an analysis of the simulations and the microphysical budget therein, physical explanations as to the reasons for such sensitivities are offered. To help narrow down the focus of the microphysical parameter space study at 100 m resolution, a larger set of 1 km simulations were first performed. Though they are not the main focus of this study, the results of these preliminary simulations are briefly presented.

The rest of this paper is organized as follows: Section 2 examines the experimental setup and the numerical modeling strategies used. Section 3 contains comparative discussion of storm evolution in seven high-resolution (100 m) model runs. Also considered are the effects of variation in microphysical parameters on specific dynamical aspects such as cold pool intensity, updraft strength, mesocyclone strength, and tornadogenesis and/or its potential. A detailed
analysis of the microphysical budget within the 100 m simulations is presented in section 4, while section 5 contains a discussion of the influence of variation in microphysical parameters on tornado formation. Finally, the results are summarized and further discussed in section 6.

2. Numerical model and experimental design

The Advanced Regional Prediction System (ARPS), Version 5, is used to perform the numerical simulations in this study. The ARPS is a compressible, non-hydrostatic NWP model suitable for storm-scale simulation and prediction (Xue et al. 2000; Xue et al. 2001; Xue et al. 2003). The most commonly used microphysics option in ARPS-based studies (e.g., Hu et al. 2006; Liu and Xue 2006; Xue and Martin 2006) is a scheme based on Lin et al. (1983, LFO83 hereafter). The LFO83 scheme is also the basis for a number of other commonly used schemes (e.g., Gilmore et al. 2004; Hong and Lim 2006). It calculates the mixing ratios of six water species; water vapor, cloud water, cloud ice, rain, snow, and hail, and deals with complex microphysical processes including the production of and the conversions among different species. In the scheme, an exponential DSD is assumed, following Marshall and Palmer (1948); this DSD contains two parameters: the slope and intercept parameters. The exponential DSD is expressed by:

\[ n_x(D) = n_{0\, x} \exp(-\lambda_x D_x), \quad (1) \]

where \( x \) denotes the hydrometeor species (rain, snow, or hail). \( n_x(D) \delta D \) in Eq. (1) represents the number of hydrometeors per unit volume of species \( x \) with diameter between \( D \) and \( D + \delta D \). \( n_{0\, x} \) is the intercept parameter for hydrometeor species \( x \). In the case of non-precipitating cloud water and cloud ice, the DSD is assumed to be monodisperse. The slope parameter, \( \lambda_x \), which is equal to the inverse of the mean size diameter of each distribution, is given by:
\[ \lambda_z = \left( \frac{\pi \rho_s n_{0x}}{\rho q_s} \right)^{1/4}, \]  

where \( q_s \) is the mixing ratio for hydrometeor species \( x \), \( \rho_s \) is the particle density, and \( \rho \) is the density of air. Single-moment microphysics schemes like that of LFO83 predict the mixing ratio of total mass that is proportional to the third moment of the DSD and specify the intercept parameter for the DSD of each species to a constant value.

Unfortunately, intercept parameters for rain, hail, and snow DSDs vary widely in nature. Intense convection typically yields a DSD that favors large raindrops, while stratiform rain usually has a DSD favoring small raindrops (Marshall and Palmer 1948). Hail DSDs are even more variable, with some storms producing only small hailstones (or none at all) and other storms producing many large hailstones of more than 10 cm in diameter. Observational studies have yielded values of rain, hail, and snow intercept parameters spanning two, four, and two orders of magnitude, respectively (Waldvogel 1974; Lo and Passarelli 1982). In typical model simulations using single-moment microphysical schemes, however, the intercept parameters are usually treated as constant, specified using “reasonable” values based on limited observations. Since a wide range of observed values exists, a better understanding on the sensitivity of model simulations to these parameters is necessary. In our current study, such simulations are for supercell storms performed at up to 100 m horizontal resolutions, and tornadogenesis and/or its potential are our focus.

In all simulations presented in this study, the storm environment in the model is defined using a horizontally homogeneous base state derived from a modified sounding for the May 20, 1977, Del City, Oklahoma storm (Ray et al. 1981). A mean storm motion vector of (3, 14) ms\(^{-1}\) is subtracted from the sounding to keep the main cell of the simulated storm near the center of
the computational domain; effectively, this computational domain follows the main storm cell. Free-slip boundary conditions are applied at the surface, and radiation boundary conditions are applied at the lateral edges of the domain. Initial convection is triggered by a thermal bubble of 4 K maximum potential temperature perturbation, with the vertical and horizontal radii being 1.5 and 5 km, respectively. Domain-wide maxima and minima of vertical vorticity and wind velocity are calculated every second during the model simulations, and 3D gridded fields are written out every 30 seconds in the 100 m simulation over the four hours of simulation time.

All existing studies that attempt to simulate tornadoes within a complete supercell storm employ nested grids, where single or multiple levels of high-resolution grids are nested within a coarser resolution grid (e.g., Grasso and Cotton 1995; Wicker and Wilhelmson 1995). This was necessary because available computational resources were insufficient to represent the entire model grid necessary to encompass the entire supercell storm at the uniform high resolution required to resolve the embedded tornado. While significant insights have been gained using such an approach through analysis of the fine grid output, the subjective introduction of nested fine grids precludes the possibility of tornadogenesis in other parts of the model domain and limits the value of the numerical simulations if one is mostly interested in when, where and if tornadogenesis would occur within the simulated supercell. The typically small size of the highest-resolution domain used also constrains one's ability to fully analyze tornadogenesis dynamics. This current study employs a large enough grid (64 km × 64 km) to fully encompass both a developing tornado and its parent supercell using a uniform horizontal resolution of 100 m. We realize that the 100 m resolution is still relatively coarse to resolve tornadoes, especially smaller ones; this choice is necessitated by the need to make multiple simulations with available computing resources.
Still, earlier numerical studies have been very valuable. Over twenty years ago, using a then-supercomputer that is less powerful than a hand-held computer today, Rotunno and Klemp (1985) analyzed simulation results created at a horizontal resolution of 500 m, and suggested that the tilting and subsequent stretching of low-level horizontal vorticity baroclinically generated along the forward flank gust front is the main source of low-level vorticity and therefore rotation in tornadoes. Some recent observational studies (e.g., Markowski et al. 2002), however, failed to find strong cold pools and associated strong gust fronts in observed tornadic thunderstorms, suggesting other processes may have been responsible for tornadogenesis. An alternative mechanism in which the downward transport of mid-level rotation associated with the mesocyclone by the rainy downdraft plays a key role is suggested by Davie-Jones et al. (2001). Both of the aforementioned studies point to the important role of the downdraft and the resultant cold pool, whose intensities are usually strongly influenced by rain evaporation processes governed by the model microphysics.

Apart from the sensitivity to microphysical parameters to be studied in this paper, other aspects of model configurations can also affect tornadogenesis and the genesis of mesocyclones. Adlerman and Droegemeier (2002) conducted a study investigating the effects of variations in ARPS model parameters, including horizontal and vertical grid resolution, numerical diffusion and drag coefficients, as well as the choice of microphysical schemes, on the cycling times of cyclic mesocyclogenesis; strong impacts from some of these parameters on cycling behaviors were found. While these past studies provide valuable insights into specific subcategories of storm behavior, none of them focuses in detail on the changes in overall supercell storm dynamics, tornado potential, and tornadogenesis, or examines systematically the impact of
Two sets of simulations were conducted; they include thirteen 1 km horizontal-resolution simulations (the low-resolution set) and seven 100 m horizontal-resolution simulations (the high-resolution set). The low-resolution set varies the snow, rain, and hail intercept parameters as well as hail density while the high-resolution set varies only rain and hail intercept parameters. Details concerning the configuration for each of the simulations can be found in Table 1. In each of the simulations, one or more intercept parameters were varied for rain, snow, or hail, along with possible variation in hail density. Control values were based on default settings in LFO83, and the increased and decreased values used in experiments were chosen to fall within the range observed values of previous studies (Waldvogel 1974; Lo and Passarelli 1982).

Our earlier simulations with the current sounding and the LFO83 ice microphysics scheme, using standard parameter settings, produced an overly intense cold pool that prevented the development of an intense tornado, while a nearly identical simulation using warm rain microphysics produced a tornado of F5 intensity when using a 25 m horizontal resolution (the results of this simulation will be the subject of a future paper). In the warm rain simulation, the cold pool was considerably weaker; runs that produced weaker cold pools than the control simulation are of particular interest here. The control run used default settings, while other experiments varied these parameters. A single simulation using warm rain microphysics, marked ‘Kessler’, is included in this study purely for purposes of comparison. The 1 km resolution simulations were run to narrow our focus within the microphysical parameter space by determining the relative magnitude of effects on general supercell dynamics, in terms of cold pool intensity and potential for tornadogenesis, introduced by varying different microphysical
parameters. Cold pool intensity is selected as an indicator of low-level dynamical sensitivity because it has a strong effect on storm propagation and on the baroclinic generation of low-level horizontal vorticity, which in turn influences tornadogenesis (Rotunno and Klemp 1985). The cold pool intensity also has a strong effect on the updraft intensity and orientation in convective systems (Rotunno et al. 1988). Having determined the microphysical parameters with greatest influence, we repeated a subset of the 1 km experiments at 100 m horizontal resolution, varying only the parameters with greater influence. The formation and development of tornadoes, if any, in the 100 m simulations, are analyzed.

For the low-resolution simulations, a larger physical domain of $128 \times 128 \times 16 \text{ km}^3$ is used, with 35 vertical levels and the vertical grid spacing increasing from 100 m near the surface to 700 m near the model top. For the high-resolution simulations, a smaller domain of $64 \times 64 \times 16 \text{ km}^3$ is used, with 81 vertical levels and the vertical grid spacing increasing from 20 m near the ground to about 400 m at the model top located at 16 km height. The reason for the difference in the horizontal domain size between the 1 km and 100 m experiments is largely due to limitations in available computational resources. When the 1 km simulations are rerun using the smaller domain used in the 100 m experiments, the results are qualitatively similar; using the larger domain allows the main cell to remain completely within the domain for a longer period of time.

3. Results

In each of the 1 km simulations, the dominant storm mode was steady or cyclic supercell. In the majority of the simulations, a single supercell forms in the vicinity of the initial warm bubble perturbation and undergoes a split between 3000 and 4000 seconds. The left mover quickly exits the domain, while the right mover remains near the center of the domain, or moves
southeast within the domain (depending on the cold pool intensity, figures not shown). Faster southeastward motion is noted in simulations with stronger cold pools. In Fig. 1 is a plot of the maximum cold pool intensity, in terms of the minimum potential temperature perturbation ($\theta'$) within the entire model domain, at the first grid level above ground, for selected 1 km runs. Fig. 1 shows a great deal of variation in the cold pool intensity among the simulations.

The greatest departure from the control run (CNTL) is obtained by altering the rain intercept parameter. The cold pool in experiment R5 (c.f., Table 1) is on average 2.62 K warmer than in CNTL over the 4 hour period, while R7 yields a cold pool on average 1.19 K cooler than in CNTL (c.f., Fig. 1a). The hail intercept parameter was also quite influential; H2 yields a cold pool on average 1.99 K warmer than in CNTL (c.f., Fig. 1b); this is consistent with the results of van den Heever and Cotton (2004), who found that decreasing the mean hail size led to a more intense cold pool in their 1-km resolution supercell simulations. They did not examine the sensitivity to rain intercept parameter, however.

The snow intercept parameter and hail density had less of an influence; S7 and S8 yield departures in the mean minimum cold pool temperature (potential temperature perturbation) of only 0.23 and 0.58 K respectively (c.f., Fig. 1c), from CNTL, while D400 yields a mere 0.10 K departure from CNTL (c.f., Fig. 1d). One possible cause for these results is that snow is largely present in the anvil and upper levels of the storm, and thus the snow DSD likely has less direct impact on low-level dynamics such as cold pool intensity and tornadogenesis. Hail density also has a much smaller impact on cold pool intensity, because the hail density has smaller direct effect on low-level evaporation rates.

The 1 km simulations show that the rain and hail incept parameters have the largest influence on the cold pool intensity and the storm dynamics in general. We therefore focus
exclusively on the effects of these two parameters in our 100 m simulations. Seven 100 m horizontal resolution experiments were conducted, varying the rain and hail intercept parameters (Table 1). They include the control simulation (CNTL), two simulations varying the rain intercept parameter (R5 and R7), two simulations varying the hail intercept parameter (H2 and H6), and two simulations varying both rain and hail intercept parameters (H2R5 and H6R7).

One hour into the simulation, the differences among the 100 m simulations are already significant, as shown by Fig. 2, which compares near-surface (10 m above ground) reflectivity, cold pool intensity, and horizontal wind fields one hour into the simulations. The cold pool is weakest (minimum surface $\theta'$ is -5.02 K) and smallest in H2R5 (Fig. 2d) and is strongest and largest in H6R7 and R7 (minimum $\theta'$ is -12.3 K and -13.6 K, respectively). The cold pools in CNTL and H2 are of similar intensity. In general, larger cold pools correspond to larger regions of reflectivity (and hence precipitation) at this time, consistent with the fact that less precipitation produces less evaporative cooling.

As more time passes, the differences in the simulation results become even greater. At the later times, the simulated storms range from a single, steady supercell, to multiple supercells, to a quasi-linear convective system among the experiments (Fig. 3 and Fig. 4). Simulations with weaker cold pools (H2, R5, H2R5) result in the formation of single or multiple supercells with steady updrafts. Supercells are also present in simulations with moderate to strong cold pools (CNTL, H6, R7), though these supercells are more cyclic in nature, with multiple updraft pulses, as shown in Fig. 5, which provides a comparison of vertical structures between a weaker cold pool case (R5) and a stronger cold pool one (H6R7). van den Heever and Cotton (2004) similarly report that decreasing the mean hail size (accomplished in our study by increasing the hail intercept parameter) in supercell simulations results in a stronger but more unsteady storm.
Unlike at 1 km resolution, in two of the simulations with the strongest cold pools (R7 and H6R7), supercells are present initially, but the system later transitions to a more linear mode (see Figs. 3f-g, 4f-g), likely due to the strong linear forcing of the intense gust front associated with the storms. The difference in organizational mode between two runs of the same simulation at different resolution serves to highlight that sensitivities are also present to the choice of model resolution, as discussed in Adlerman and Droegemeier (2002).

Simulations DSDs favoring larger hydrometeors (H2, R5, and H2R5) correspond to the weakest cold pools. Combining larger raindrops and larger hailstones seems to have an additive effect, with H2R5 exhibiting the weakest cold pool of all. The opposite is true for simulations with DSDs favoring smaller hydrometeors (H6, R7, and H6R7); in these simulations the cold pool is more intense than in CNTL and experiments favoring larger hydrometeors than CNTL. Again, combining smaller raindrops with smaller hailstones has an additive effect, with the most intense cold pool of all observed in H6R7. One of the main causes of these results is the differences in evaporative cooling among different simulations, as will be demonstrated in the microphysical budget analysis.

Fig. 6 shows the rainwater mixing ratio fields near the surface for all seven 100 m runs at 3 hours of model time; by this time the storms had matured, and tornadic activity is present in CNTL and R5. Figs. 6b-6d show that although the maximum rainwater mixing ratio in simulations favoring larger hydrometeors is higher than that of the simulations favoring smaller hydrometeors, the rain is confined to a smaller area, limiting the spatial extent of the evaporative cooling. The result of these effects is a smaller, weaker cold pool compared to CNTL. The opposite effect is at work in simulations with smaller DSDs, which result in more total hydrometeor surface areas and wider spatial distributions (Fig. 6e-g), allowing for more
evaporation and thus more cooling of the air in the cold pool than in CNTL, resulting in a larger, stronger cold pool. The presence of tornadic activity in CNTL and R5, in which weak to moderate cold pools with relatively limited area coverage were noted, reinforces the observational findings of Markowski et al. (2002), whose mobile mesonet observations demonstrated that “relatively cold, stable surface air parcels were found to be more widespread in nontornadic RFDs”, as well as their conclusion that “tornado likelihood, intensity, and longevity increase as the surface buoyancy, potential buoyancy (CAPE), and equivalent potential temperature in the RFD increase”.

In the case of variation of the rain intercept parameter, differences in evaporative cooling result largely from the difference in total droplet surface area. In those simulations with DSDs favoring larger rain drops, for the same total volume of rain water, the total surface area for rain drops is smaller therefore the rain evaporation is less as the rain drops fall. As a result, more evaporation can take place when the DSD favors small raindrops, enhancing the evaporative cooling that results. Similar reasoning is also true for hailstones. In addition, smaller droplets are lighter and have a lower terminal velocity; they are more readily advected away from the main updraft and fall out over a wider area. The result is a larger spatial distribution of evaporative cooling, and thus a cold pool with more area coverage. These conclusions are supported by Fig. 4 and Fig. 6, which show that in R5 areas of light rain are diminished but the cores of intense rainfall exist because the larger, heavier raindrops with their larger terminal velocities are not advected as far downwind before falling to the surface. As seen in Fig. 4, the cold pool in R5 is relatively weak compared to that in CNTL, and is less extensive in area coverage. By contrast, R7 has larger areas of lighter rain as a result of smaller, lighter droplets.
being advected farther downwind before falling to the surface; an enhanced cold pool with larger area coverage therefore forms as a result of the enhanced evaporative cooling.

By contrast, variation in the hail intercept parameter has only a weak direct effect on evaporative cooling, but a stronger influence on the distribution and intensity of rainfall. As seen in Fig. 6, modifying the hail DSD to favor larger hailstones results in a notable decrease in the area coverage of rainfall and an increase in localized intense rain cores. The effect is similar to, but more marked than, that seen in R5. The opposite effect can be seen in H6, where shifting the hail DSD to favor smaller hailstones results in a reduction in the intensity of the rain cores, but also induces a dramatic increase in the total area coverage of rainfall. As a result of these effects, as seen in Fig. 4, H2 produces a weaker, more confined cold pool than CNTL, and H6 produces a slightly stronger cold pool with greater area coverage. These results are consistent with those of van den Heever and Cotton (2004) in that a hail DSD favoring small hailstones resulted in a stronger cold pool covering a larger geographical area. The process is similar to the rain case, as smaller hailstones are advected farther, when they melt to form rain drops, the rainfall coverage is increased.

4. Microphysical Budget Analysis

The importance of the differences in the contributions from different microphysical processes between the 100 m simulations, and the dominant role of evaporation in producing these differences, can best be demonstrated by detailed analysis of the microphysical and thermodynamic budgets in these simulations. We will consider the relative magnitude of the heat sources and sinks provided by each of the relevant microphysical processes, and the differences in the role of the dominant microphysical processes among the simulations. To
perform microphysics-related budget analysis, codes were developed and added to the ARPS that calculate the total mass of water converted from one species to another at each timestep within a specified region of the model domain for each of the more than two dozen microphysical conversion terms within the Lin scheme. This value is then multiplied by the time step and the relevant latent heat (of fusion, vaporization, or sublimation) for the process to obtain a total heating contribution from each microphysical conversion term.

The primary goal of the budget analysis is to compare the relative magnitudes of cooling contributions of microphysical processes to cold pool intensity. Thus, to account for the cooling related to the developing cold pool, the budget calculations are limited to the region from the surface to 5 km height and where the vertical velocity is less than - 0.5 m s\(^{-1}\). The calculations are further limited to between 3600 and 7200 s, a time during which the cold pool of the initial supercell develops to maturity. Microphysical cooling contributions are output once every 3 minutes in the one hour period. The time series of the dominant cooling terms, i.e., those of rain evaporation and graupel melting, are shown in Fig. 7.

Within the developing cold pool, two microphysical processes are found to be dominant in terms of their cooling contribution. The largest contribution comes from the evaporation of rain, with a secondary contribution from melting of graupel (not shown). The ratio of the cooling contribution of melting of graupel to that of evaporation of rain ranges from approximately 0.1 in H6R7 to approximately 0.55 in R5. All other microphysical conversion terms were found to be negligible in their cooling contribution compared to evaporation of rain and melting of graupel (not shown).

There are large differences in the amount of cooling in downdraft regions below 5 km due to evaporation of rain. When the total cooling is integrated from 3600 to 7200 s and
normalized as a fraction of cooling observed in control run CNTL (see Fig. 7a), total cooling contributions from evaporation of rain range from 0.25 in H2R5 to 2.28 in H6R7. Significantly less cooling from evaporation of rain is observed in H2 and R5, which have 0.53 and 0.47 times the cooling observed in CNTL, respectively. Conversely, H6 and R7 show significantly stronger cooling contributions from evaporation of rain than CNTL, by ratios of 1.54 and 1.59 respectively. As discussed earlier, the differences from CNTL in H2 and H6 are largely due to differences in spatial coverage of the cold pool (see Fig. 6), while in R5 and R7 the differences are largely due to the difference in the raindrop DSD hence the total raindrop surface area that affects rain evaporation.

Variation of the cooling contribution from melting of graupel is much less pronounced between the runs. Total cooling contributions integrated between 3600 and 7200 s range between 0.82 and 0.9 of that observed in CNTL for R5, R7, H6, and H6R7 (see Fig. 7b). Significantly less cooling from melting graupel was observed in H2 and H2R5, which had cooling contributions of 0.48 and 0.35 of that seen in CNTL respectively. Relatively little change in cooling due to melting of graupel was introduced by varying the rain DSD, as evidenced by the similar curves for R5 and R7 in Fig. 7b. The only significant departure from CNTL was obtained in the two runs using a DSD favoring larger hailstones (H2 and H2R5). In these runs, the total hail/graupel surface area should be much less than in other runs (because of fewer larger hailstones), leading to less total melting. This process is similar to that seen in raindrops leading to less total evaporation of rain in R5 and H2R5. From the above budget analysis, we can conclude that the biggest source of difference in cold pool intensity among the 100 m simulations lies with the evaporative cooling processes associated with rain water.
5. Tornado formation and development

Sustained tornadic low-level vortices are observed in two of seven 100 m simulations: CNTL and R5, as noted in Fig. 8. The figure shows the time series of maximum low-level vertical cyclonic vorticity from different 100 m simulations, with prominent tornadic circulations noted. Tornadic circulations appear as extended peaks in the vorticity time series data. The low-level vortex in CNTL lasts for approximately 4 minutes, with a maximum low-level wind speed of 55 m s\(^{-1}\), giving the vortex an F2 intensity on the Fujita scale. The near surface vortex in R5 lasts approximately 10 minutes, with a maximum surface wind speed of 58 m s\(^{-1}\) (also F2 intensity). Both CNTL and R5 simulations have well-defined supercells and weak to moderate cold pool intensities. In contrast, surface vortices that form in simulations with the strong cold pool are all weak and short lived. Brief, weak vortices are observed in several simulations that usually last less than 90 seconds. These include a series of five spinups in CNTL between 8000 s and 10000 s, as shown in Fig. 8b, as well as two spinup attempts in R5 between 8600 s and 9400 s, each lasting approximately one minute. In both CNTL and R5, these spinups occur beneath the mesocyclone of the northern supercell storm in the domain (not shown), in the same storm-relative location as the later long-lived tornado. In other simulations, these spinups occur along the gust front, a likely result of baroclinically generated crosswise horizontal vorticity being tilted into the vertical and stretched (not shown). We will focus in this study on the longer-lasting tornadic vortices in CNTL and R5 that are actually associated with supercell mesocyclones, rather than on the weak, short-lived ones.

In CNTL, the strongest tornado occurs between 12300 s and 12600 s, and is characterized by a vertical vorticity maximum of between 0.5 and 0.75 s\(^{-1}\). The tornado in R5 occurs between 12000 and 12780 s, with vertical vorticity maxima also between 0.5 and 0.75 s\(^{-1}\).
From initial time to 1800 s, the vorticity time series are nearly identical in all runs. After this, maximum vorticity increased from an average value of just under 0.25 s\(^{-1}\) to between 0.25 s\(^{-1}\) and 0.40 s\(^{-1}\) (Fig. 8). Lower average values of maximum vorticity are seen in H6R7 (a case with the strongest cold pool and least supercellular storm mode) and H2 (a case with cold pool intensity similar to that of the control run).

The effect of the cold pool intensity on the storm dynamics and tornadogenesis can be, to an extent, revealed by the vertical cross-sections through the updrafts of convective cores, for experiments R5 and H6R7 that have weak and strong cold pools, respectively (Fig. 5). These cross-sections show that the positioning and strength of the gust front are strong factors in controlling the orientation of the updraft. In simulations with weaker cold pools, the gust front is positioned beneath or just ahead of the mid-level updraft core, promoting a vertically oriented intense updraft. In simulations with very strong cold pools, including H6R7, the gust front quickly races ahead of the mid-level updraft core and (potential) mesocyclone circulation, often by several kilometers (see Figs. 2, 3, 9). The result is a gentler, slanted updraft, often with multiple updraft “pulses”, promoting cyclic behavior or in extreme cases linear rather than supercellular storm mode.

These differences in vertical structure can also be directly deduced from Fig. 9, which show overlays of surface cold pool intensity and mid-level (2.5 km) wind vectors. From these overlays we see that mid-level circulation centers are positioned farther behind the gust fronts in cases such as R7 and H6R7 where the cold pool was very intense, while the circulation centers are positioned much closer to the gust front, allowing any surface circulations present to benefit from enhanced vortex stretching due to better vertical alignment of low- and mid-level vertical vorticity centers.
Figure 10 shows the trajectories of the air parcels originating from the low-level inflow region ahead of the forward flank gust front, in R5 and H6R7. Parcels were released into the inflow at such time that they would enter a mature cell near its peak intensity. In R5, the inflow parcels begin near the surface, and are drawn southwestwards relative to the storm, remaining near the surface until entering the updraft and rising with a moderate westward slope to the mid-levels, followed by a nearly vertical ascent to the upper levels of the storm. Some of the parcels reach altitudes of nearly 14 km before sinking, indicating a storm with vigorous updraft and an overshooting top. While ascending from near the surface to mid levels (4 km altitude) the parcels moved an average of 2 km horizontally, for a resulting steep updraft slope of 2 over 1. The parcels within the inflow region of the storm in H6R7 similarly began by moving southwestwards near the surface, but upon entering the updraft they maintained a strong westward component in their trajectory as they rose to the mid-levels, with somewhat more vertical motion when entering the upper part of the updraft. The parcels in the inflow of H6R7 move an average of 6 km horizontally while they rise from near the surface to the mid-levels (4 km altitude), giving an updraft slope of 0.67. The results of this trajectory analysis further highlight the difference in the updraft structure, which has implications on the low level vertical stretching, an important process in low-level vortex intensification, as discussed above. The much larger rearward velocity component of the parcel trajectory can be explained by the stronger baroclinic generation of horizontal vorticity along the gust front in H6R7 (Rotunno et al. 1988).

a. Tornadoes in control experiment CNTL

In control experiment CNTL, two tornadic events occur: between 8400 and 9600 s, a series of very brief, short-lived tornadic spinups takes place (Fig. 8). These spinups are all very
brief, lasting no more than two minutes each, and the maximum wind speed in these spinups never exceeds 50 m s$^{-1}$, putting these short-lived tornadoes within the F0 to F1 range on the Fujita scale. A more long-lived tornadic vortex of F2 intensity is present between 12300 and 12600 s, lasting between 4 and 5 minutes, and achieving a maximum wind speed of approximately 55 m s$^{-1}$ (not shown). At the time of these tornadoes, two supercell thunderstorms are present within the model domain, a dominant southern storm, which formed around 6000 s along the rear-flank gust from of the initial storm and attained supercell characteristics by 7500 s. The other is the initially dominant but later decaying northern storm that is the right-mover of the original storm that developed from the initial bubble. This storm exhibits supercell characteristics from 3000 s to 13000 s (Fig. 2a, Fig. 3a and Fig. 4a). All tornadic activity in this simulation is associated with this northern storm.

The first spinup attempt in CNTL, occurring between 8100 s and 8220 s, is located beneath the mid-level mesocyclone, near the intersection of the storm’s rear and forward flank gust fronts. At this time, the forward flank gust front is somewhat diffused and poorly defined (not shown). Subsequent spinups follow within several minutes, between 8500 and 9200 s, similarly situated beneath the mesocyclone near the intersection of the rear and forward flank gust fronts, within a hook echo present in the radar reflectivity field. None of these individual vortices last longer than two minutes. Horizontal velocities within these vortices range from 40 m s$^{-1}$ to 50 m s$^{-1}$ (F0 to F1 on the Fujita scale).

In the above case, the individual vortices initially form along the forward flank gust front, and intensify as they are advected into the region of the hook echo, aided by vertical stretching they experience beneath the mesocyclone. Maximum intensity is reached as the vortices begin to be advected eastward by rear-flank winds, and rapid weakening follows as the vortices become
occluded, wrapping themselves in rain-cooled air from behind the rear flank gust front. Such processes are close to that described by Rotunno and Klemp (1985) in which generation of horizontal vorticity along the forward flank gust front and subsequent tilting are believed to be important.

Between 10100 s and 10500 s, a single tornadic vortex forms in CNTL (c.f., Fig. 8) in the hook echo region of the northern supercell (not shown). By this time the northern storm is beginning to decay. In contrast to the earlier series of small vortices, the tornado that develops at this later stage is a single vortex rather than one of many small vortices, and though it occurs beneath the parent mesocyclone, the tornado forms several kilometers behind the gust front, within the cold pool of the storm. During the 4 to 5 minutes that the tornadic vortex is present at the surface, it reaches a maximum wind velocity of 55 m s\(^{-1}\). The genesis process of this tornado is likely different from the earlier ones but a detailed analysis of the genesis mechanism is beyond the scope of this paper.

b. Tornadogenesis in experiment R5

In experiment R5, which favors larger raindrops, a single intense tornado is present approximately between 12,000 s and 12,720 s, lasting for about 10 minutes and attaining a maximum wind speed of 58 m s\(^{-1}\) (F2). Like the control run, two supercell thunderstorms are present within the model domain, and all tornadic activity is associated with the northern storm. At 11,940 s, though the southern storm has become the dominant storm within the domain, the northern storm remains a mature supercell, with a well-defined mesocyclone and a clearly evident hook echo in the radar reflectivity field near the surface. The tornadic circulation begins to develop near the tip of the hook echo, beneath the mid-level mesocyclone, and at the intersection of the rear flank and forward flank gust fronts. Unlike in CNTL, the forward flank
gust front in R5 is much better defined. The tornado reaches its maximum intensity about five minutes later at 12,360 s, and the low-level tornadic circulation attains a diameter of approximately 700 m, or 7 times the grid spacing. The tornado persists at near maximum intensity for about 2 minutes, until around 12,480 s, and then begins a steady weakening trend until its end around 12,720 s. Plots of the reflectivity, vorticity, and surface wind for this tornado, near its peak intensity, are shown in Fig. 11. The tornadic circulation, embedded within the hook echo of its parent supercell is clearly visible.

As with the spinups early in CNTL, the weakening of the circulation is brought about as cold air from within the storm’s cold pool is drawn all the way around the tornado by its circulation, making the low level flow less buoyant. Unlike in CNTL, however, the cold pool in R5 is relatively weak, which seems to allow the tornado to remain strong for a longer period of time, with an intense surface vortex discernable for around 10 minutes, as opposed to 4 minutes in CNTL. In addition, the tornado in R5 is located much closer horizontally to the associated mesocyclone than is the case in CNTL. A plot detailing mesocyclone and tornado locations at the time of peak tornado intensity in both CNTL and R5 is shown in Fig. 12. In CNTL, the tornado, marked with a “T”, is located approximately 8 km east of the center of circulation of its parent mesocyclone, marked with an “M”. The tornado in R5 is located only about 3 km northeast of the center of circulation of its parent mesocyclone. Because the tornado in R5 is located almost directly beneath its parent mesocyclone, it probably benefits more from the mesocyclone-induced dynamic (upward pressure gradient force, Rotunno and Klemp 1982; Klemp 1987) forcing than that in CNTL, which promotes a stronger updraft and more vertical vortex stretching.
The greater longevity of the tornadic circulation in R5, where cold pool intensity was relatively weak, once again provides modeling evidence to reinforce the observation-based conclusions of Markowski et al. (2002) that tornado longevity is increased in environments with warmer, more buoyant near-surface parcels in the RFD region. From these results, we can surmise that the microphysical parameters chosen to represent rain and hail DSD have a strong influence on tornadogenesis potential through their effects on cold pool intensity and coverage. Tornadogenesis potential is greater when there is a near-vertical alignment of the low and mid-level vorticity centers, and when the air in downdraft region is warmer and thus more buoyant. Changing the DSD to favor larger raindrops (as in R5) results in a decrease in evaporative cooling, leading to a weaker cold pool that does not propagate very far ahead of its parent storm. This not only provides relatively warm, buoyant air in the downdraft region, but leads to a more vertical updraft orientation as well, ensuring that the gust front is positioned near or beneath the parent mesocyclone, providing additional tornadogenesis potential due to dynamic forcing (Klemp 1987). The opposite is true in cases where the DSD favors smaller raindrops (as in R7), where tornadogenesis potential is greatly limited by enhanced evaporative cooling, causing a strong cold pool that quickly propagates well ahead of its parent storm, leading to a gentler, slanted updraft that separates the low and mid-level vorticity centers, as well as very cold, less buoyant air within the downdraft region. Similar arguments can be applied to changing the hail DSD to favor larger hailstones (as in H2), a change which increases tornadogenesis potential by leading to a decrease in precipitation coverage, causing a less extensive cold pool. Conversely, changing the hail DSD to favor smaller hailstones (as in H6) decreases tornadogenesis potential by increasing precipitation coverage, leading to a more extensive cold pool and more slanted updrafts. We should point out here, though, that the condition about the alignment of lower and
middle level circulations and the updraft orientation is more general than having larger or smaller raindrops or hailstones. The latter can be dependent on the environmental thermodynamic and wind conditions.

c. **Pressure perturbations within tornadoes**

Fig. 13 shows a time series of domain-wide perturbation pressure during the period of most intense tornadoes in CNTL and R5. During the period that the tornadoes are on the ground, the largest negative pressure perturbation is near the surface, collocated with the tornado, not surprisingly due to the centrifugal force associated with the strong tight vortex. In CNTL, a maximum pressure deficit is 21.5 hPa, and in R5 it is 23 hPa. These pressure deficits are less than the 50 to 60 hPa that Winn et al. (1999) observed in the vicinity of the F4 tornado of June 8, 1995 near Allison, Texas. These results are consistent, however, when compared with the two-way nested grid simulation of Grasso and Cotton (1995), who observed a maximum pressure deficit of 29 hPa in a tornado simulation using a similar sounding based on the May 20, 1977 Del City storm and a horizontal resolution of 111 meters on a nested grid. The pressure deficits in these simulations are less than some of the observed values associated with more intense tornadoes can be due to the still low 100 m resolution which can limit the size and intensity of the simulated tornado and therefore result in a broader, weaker vortex with weaker pressure perturbation. To more accurately resolve a tight intense tornado as well as its internal structures would likely require a horizontal resolution no coarser than 10 m (Davies-Jones et al. 2001).

Further, trajectories of parcels entering the tornado vortex near the ground were also analyzed, and the results of these analyses (not shown) were consistent with other very high resolution (up to 12.5 m) tornado simulations conducted by the second author. Since a detailed analysis of the tornadogenesis mechanism requires much more space than available in this paper,
and the main goal of this paper is the investigation of the sensitivity of tornadogenesis to microphysics, we will not go any further with respect to the analysis of tornadogenesis mechanisms. Such detailed analyses will be reported elsewhere.

6. Summary and further discussion

In this study we have demonstrated that numerical simulations of supercell thunderstorms are highly sensitive to drop size distribution parameters in a commonly-used single-moment ice microphysic scheme, both at horizontal resolutions of 1 km and 100 meter. Using the Lin et al. (1983) ice microphysics scheme and varying only the intercept parameter for the Marshall-Palmer-like exponential drop size distributions of rain and hail, we obtained widely varying model solutions, including single and multiple supercells and a linear system, from the same set of initial and environmental conditions. We have shown that cold pool intensity and geographical coverage, precipitation intensity and amount, updraft speed and orientation, and tornadogenesis all have significant sensitivity to variations in rain and hail intercept parameters. The sensitivities of the simulated supercell storm to microphysical parameters observed in this study are generally consistent with the findings of Gilmore et al. (2004) and van den Heever and Cotton (2004), whose simulations are limited to the 1 km horizontal resolution, a resolution insufficient to explicitly resolve tornadogenesis.

Simulations at the 1 km horizontal resolution reveal that, of the parameters tested, cold pool intensity and low-level supercell dynamics are most sensitive to variations in the rain and hail intercept parameters and much less sensitive to the snow intercept parameter and hail density. While rain and hail are present in high concentrations near the surface and at the mid-levels, snow in this case is present only in the upper levels and in the anvil region of the storm, lessening its impact on low-level dynamics and properties such as cold pool strength and
tornadogenesis potential. Also, the snow intercept parameter and hail density have much less influence on the amount of evaporative cooling present in the unsaturated air beneath the storm, a major factor in determining cold pool strength and the resulting storm dynamics.

In our 100 m simulations, we demonstrated that when the DSD used favors larger hydrometeors (smaller intercept parameters), the resulting simulations exhibit weaker cold pools. Reduced total hydrometeor surface area associated with larger raindrops or hailstones leads to less evaporation of hydrometeors passing through unsaturated air, reducing total cooling due to evaporation or melting. In addition, the larger hydrometeors, with their greater terminal velocity, are not advected very far from the updraft before falling to the surface, reducing the geographical coverage of the precipitation core, further limiting the cold pool intensity. The weaker cold pool, demonstrated in our budget analysis to be due primarily to reduced evaporative cooling, produces supercell storms with steady, strong updrafts, and high tornadogenesis potential due to a favorable gust front that allows for vertically oriented updrafts and better alignment of low-level and mid-level vorticity centers. By contrast, when the drop size distribution favors smaller hydrometeors (larger intercept parameters), the resulting simulations contain many smaller hydrometeors, with a much greater total surface area for evaporation, and a lower terminal velocity. Thus the storms produced exhibit very intense cold pools due to enhanced evaporative cooling over a larger geographical area, resulting in cyclic supercells or quasi-linear convective systems aligned along the intense, surging gust front. These storms tend to have weaker, pulsing updrafts that slant rearward with height, due to the strong baroclinically generated vertical circulation at the gust front (c.f., Rotunno et al. 1988). The rearward slanting updraft positions the mid-level mesocyclone farther from the low level vortices that typically form close to the gust front, reducing the potential of dynamic suction effect by the mesocyclone (Rotunno and
Klemp 1982) and the associated low-level vertical stretching. Further, a weaker, rear-ward slanting updraft by itself is ineffective in providing strong vertical stretching that is important for vortex intensification. Therefore, in the cases where the gust front is so strong that it outruns its parent storm by a large distance (by several to tens of kilometers), tornadogenesis potential is low. For these reasons, tornadogenesis is sensitive to microphysics, mainly through its impact on cold pool intensity and gust front position, in the cases studies in this paper at least.

Finally, we note that while we believe the conclusions drawn in this paper about the very large sensitivity of thunderstorm dynamics and tornadogenesis and/or its potential to microphysical parameters and the mechanism of sensitivity are robust, tornadoes do not necessarily form when the cold pool is colder or warmer. There is probably a balance between the cold pool strength and environmental flow, and the storm dynamics will certainly also be influenced by other parameters of the storm environment, including convective available potential energy (CAPE). Intense supercell tornadoes would only occur when many conditions are favorably met, including the presence of an intense rotating updraft that is located almost directly over the low-level center of rotation. Full understanding of tornadogenesis will require much more research and more detailed diagnostic analysis of high-resolution simulation as well as observational data; these can be topics for future studies.

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Table 1. Summary of experiments and their horizontal resolutions

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