Sensitivity of Real-data Simulations of the 3 May 1999 Oklahoma City Tornadic Supercell and Associated Tornadoes to Multi-moment Microphysics. Part I: Storm- and Tornado-scale Numerical Forecasts

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Abstract

Numerical predictions of the 3 May 1999 Oklahoma City, Oklahoma tornadic supercell are performed in a real-data framework utilizing telescoping nested grids of 3-km, 1-km, and 250-m horizontal grid spacing. Radar reflectivity and radial velocity from the Oklahoma City WSR-88D radar are assimilated using a cloud analysis procedure coupled with a cycled 3DVAR system to analyze storms on the 1-km grid for subsequent forecast periods. Single-, double- and triple-moment configurations of a multi-moment bulk microphysics scheme are used in several experiments on the 1-km and 250-m grids to assess the impact of varying the complexity of the microphysics scheme on the storm structure, behavior, and tornadic activity (on the 250-m grid). This appears to be the first study of its type to investigate single- vs. multi-moment microphysics in a real-data context.

It is found that the triple-moment scheme overall performs the best, producing the smallest track errors for the mesocyclone on the 1-km grid, and stronger and longer-lived tornado-like vortices (TLVs) on the 250-m grid, closest to the observed tornado. In contrast, the single-moment scheme with the default Marshall-Palmer rain intercept parameter performs poorly, producing a cold pool that is too strong, and only weak and short-lived TLVs. The results in the context of differences in latent cooling from evaporation and melting between the schemes, as well as implications for numerical prediction of tornadoes, are discussed. More generally, the feedbacks to storm thermodynamics and dynamics from increasing the prognostic detail of the hydrometeor size distributions is found to be important for improving simulation and prediction of tornadic thunderstorms.
1. Introduction

The parameterization of cloud and precipitation microphysics (MP) processes within numerical models remains one of the most important and challenging issues for accurate simulation and prediction of deep convective storms. In this study we restrict our discussion to bulk MP parameterization schemes (BMPs), which typically assume an underlying functional form (but see Kogan and Belochitski 2012) for the drop or particle size distribution (DSD or PSD) and predict one or more moments of that distribution for various hydrometeor categories. BMPs are the type of MP scheme used in numerical weather prediction models, and in most modeling studies of convective storms. Numerous studies (e.g., McCumber et al. 1991; Ferrier et al. 1995; Gilmore et al. 2004; van den Heever and Cotton 2004; Cohen and McCaul 2006; Milbrandt and Yau 2006a, b; Lerach et al. 2008; Snook and Xue 2008; Morrison et al. 2009; Dawson et al. 2010; Jung et al. 2010; Van Weverberg et al. 2010; Morrison and Milbrandt 2011; Van Weverberg et al. 2011; Jung et al. 2012; Lerach and Cotton 2012; Morrison et al. 2012; Van Weverberg et al. 2012; Van Weverberg 2013) have shown that the BMP is a substantial source of model uncertainty on the convective scales for metrics such as surface precipitation type and amount, storm morphology and propagation, convective downdrafts and cold pools, and even tornadogenesis potential.

The characteristics of the cold pool associated with the rear-flank downdraft (RFD) of supercells have been found by observational and idealized numerical studies to be a significant factor impacting tornadogenesis (Leslie and Smith 1978; Markowski 2002; Markowski et al. 2002; Markowski et al. 2003; Lerach et al. 2008; Snook and Xue 2008; Lerach and Cotton 2012). In contrast, studies (e.g., Trapp 1999; Markowski et al. 2011) have also found that the existence and strength of the mid- and low-level mesocyclone is not necessarily a strong predictor of
tornadoes. However, Trapp et al. (2005) showed that low-level mesocyclones (bases < 1 km AGL) were much more often tornadic (40%) than mid-level mesocyclones (bases 3-5 km AGL; 15%). Snook and Xue (2008) found that the presence or absence, duration, and intensity of simulated tornado-like vortices (TLVs) within simulated supercells were strongly dependent on the strength of the cold pool. In particular, they examined the sensitivity of TLVs to the choice of the (fixed) intercept parameter $N_0$ for the assumed exponential distributions of rain and hail in a single-moment (1M) BMP scheme. Smaller values of $N_0$ that favored relatively large rain drops and hailstones resulted in 1) weaker cold pools, 2) a more vertically stacked low-to-mid-level updraft and mesocyclone (better vertically aligned cold pool gust front and mesocyclone), and 3) stronger and longer-lived TLVs relative to larger values of $N_0$ that favored smaller raindrops and hailstones. These characteristics were due to overall smaller evaporation and melting rates in the smaller-$N_0$ simulations. Similarly, Lerach et al. (2008) investigated the role of initial cloud condensation nuclei (CCN) concentrations on simulated supercell tornadogenesis, finding that more polluted environments (i.e., higher CCN concentrations) favored more intense and longer-lived TLVs. Their results were also attributed to weaker cold pools and more vertically stacked updrafts and mesocyclones, again due to the overall larger precipitation particles produced. Lerach and Cotton (2012) compared the CCN effect with that of varying the low-level moisture profile and found that changes in the low-level moisture profile had the greatest effect on tornadogenesis, although the CCN effect was still evident. They found that higher moisture content in the low levels produced stronger updrafts (owing to the increased CAPE as the temperature profile remained fixed), more precipitation, and a net increase in evaporative cooling in the downdrafts despite the lower evaporation potential in the moister low levels, reducing the overall tornado potential.
Dawson et al. (2010, hereafter D10) found that the multi-moment MP scheme of Milbrandt and Yau (Milbrandt and Yau 2005a, b; 2006a; b, hereafter MY05a,b; MY06a,b, respectively) was able to simulate better the cold pool and reflectivity characteristics of the 3 May 1999 central Oklahoma tornadic supercell storms. The Milbrandt and Yau BMP package contains 1M, double-moment (2M) and triple-moment (3M) versions of the scheme, and were all examined in D10. We will hereafter refer to the versions as the MY# schemes, where “#” indicates the number of moments (1, 2, or 3) predicted. In particular, D10 found that smaller and weaker cold pools were produced with the 2M or 3M version of the MY scheme, which were more consistent with fixed and mobile surface mesonet observations of the storms on that day (Markowski 2002). The D10 study attributed the improvement to more physically realistic microphysical processes (such as gravitational size sorting) when more realistic PSDs of hydrometeors are predicted by the multi-moment schemes. The results were consistent with those of MY06b. D10 was also one of the first studies that examined the effects of a 3M-BMP scheme on the simulation of tornadic thunderstorms, though they did not investigate the impact on tornadogenesis and behavior.

Almost all of the aforementioned studies employed idealized frameworks where the storm environment was assumed horizontally homogeneous and initialized by a single sounding while the storm itself was triggered by an artificial thermal bubble (the exception was MY06a,b who investigated a large hail-producing supercell in a real-data framework). Potentially important physical processes including surface processes and radiation effects are usually excluded in such a framework, as are the effects of environmental inhomogeneity. For example surface friction has recently been shown to have a strong—even critical—impact on tornadogenesis (Schenkman et al. 2014), while the radiative effects of anvil shading have
complex impacts on the propagation of storm outflow via modification of the near storm wind profile (Frame and Markowski 2010, 2013). While such studies are important for elucidating the basic physical and dynamic processes that are important for different modes of convection, they have limitations as far as the prediction of real atmospheric convection is concerned. To more faithfully represent the behavior of real atmospheric convection, studying such convection within ‘real-data’ frameworks is very important. Working within a real-data framework also allows us to compare directly the simulations with observations; this is the approach taken in this study.

Only a few studies have attempted to predict real tornadoes or TLVs. Mashiko et al. (2009) and Schenkman et al. (2012) are two such studies. Mashiko et al. (2009) simulated convective storms in the outermost rainband of a land-falling typhoon that exhibited the characteristics of a mini-supercell, with one of the simulated storms spawning a tornado. No direct comparison, however, of the simulated tornado was made with the actual tornadoes. In Schenkman et al. (2012), a mesoscale convective system (MCS) was initialized by assimilating radar and other high-resolution observations on a 400 m grid. A TLV corresponding to an observed tornado was rather accurately simulated on a further nested 100 m grid; the tornadogenesis processes were analyzed in detail. Xue et al. (2014) documented a successful simulation of supercell tornadoes using 50-m grid spacing and Schenkman et al. (2014) performed detailed diagnostic analyses of the source of vorticity feeding the tornado vortex. None of these studies, however, examined the sensitivity of simulations to the BMP, and all used the more or less standard Lin et al. (1983) type 1M-BMP scheme\(^1\).

\(^1\) Schenkman et al. (2012) did use a reduced intercept parameter for rain following Snook and Xue (2008) within the Lin scheme to produce more realistic cold pools.
This particular work extends the study of D10 for the 3 May 1999 Oklahoma tornadic supercell case by employing a much more realistic framework, in which the initial conditions of the simulations are obtained through the assimilation of frequent Doppler radar data as well as other high-resolution observations. The simulations utilize a relatively complete physics package, including boundary layer and surface physics, subgrid-scale turbulence, radiation, and MP. Telescoping nested grids with realistic terrain are used to achieve sufficiently high resolution for simulating the TLVs. Compared to D10 which focused on examining the sensitivity of simulated cold pool strength and general structure of simulated reflectivity to the BMP configuration in a typical idealized, horizontally-homogeneous environmental setting, this current study examines and explains the impact of multi-moment vs. 1M-BMPs for the numerical prediction of the 3 May 1999 Oklahoma City tornadic supercell thunderstorm and TLVs. We first examine the behavior of the storms simulated at 1-km grid spacing (which is nested within a 3-km grid) during the forecast period when employing several variations of the MY scheme. We then go one step further by performing simulations with a 250-m grid spacing, nested inside the 1-km grid, so as to assess the effects of the BMP on the prediction of TLVs within the simulated supercell, since simulating tornado processes require much higher resolutions than typically needed for supercell simulations.

As Part I of a two-part paper series, we focus in this paper on the sensitivity of various aspects of the tornado and parent supercell storm simulations to the BMPs used, and in particular how the number of moments predicted affects the simulation results. Part II will focus on the dynamical effects related to tornado behavior and in particular the forces that cause the rapid vertical acceleration of flow above the ground and the intensification of the tornado vortex through vertical stretching. The organization of the rest of this paper is as follows. In section 2,
we describe the setup of the data assimilation and forecast experiments. Section 3 describes the results of the 1-km grid experiments. Comparisons of the predicted track of the simulated mesocyclone with the observed mesocyclone are made, as well as comparisons of the simulated cold pool with Oklahoma Mesonet (Brock et al. 1995) observations. Additionally, the latent cooling budget in the downdrafts due to microphysical processes is analyzed in detail. Section 4 describes the results of the innermost nested grid of 250-m grid spacing, focusing on the simulated TLV tracks and their comparison with the observed track. Various aspects of the TLVs, such as their duration and intensity in relation to the cold pool evolution are also discussed. Section 5 summarizes the paper.

2. Experiment methodology and event overview

a. Overview

Similar to the studies of Schenkman et al. (2011; 2012) and Xue et al. (2014), we use the ARPS model (Xue et al. 2000; 2001; 2003) for forward prediction during the data assimilation cycles and for the ensuing forecasts, and use the ARPS 3DVAR (Gao et al. 2004; Hu et al. 2006b) and its complex cloud analysis system (Zhang et al. 1998; Xue et al. 2003; Hu et al. 2006a) for data assimilation. Subgrid-scale turbulence is predicted using 1.5-order TKE and PBL parameterizations, the latter based on Sun and Chang (1986). Radiation physics are based on the NASA Goddard long- and short-wave parameterization scheme (Chou 1990, 1992; Chou and Suarez 1994). A two-layer land surface model described in Xue et al. (2001) is employed. We use three levels of one-way nested grids with grid spacings of 3 km, 1 km, and 250 m, respectively. These grids and associated dimensions are shown in Fig. 1. The grids are designed to take advantage of different data sources in such a way as to capture the mesoscale storm environment on the 3-km grid, and the convective storms themselves via radar data
assimilation on the 1-km grid. Experiments on the 250-m grid are intended to capture the internal structures and circulations of the convective storms, including near-tornado-scale features. No data assimilation is performed on the 250-m grid. Vortices that form in the hook echo and RFD region of the simulated storms that achieve tornado-like magnitudes (when computed on the scale of a typical tornado) of vertical vorticity $\zeta \sim O(0.1-1.0 \, s^{-1})$ and horizontal wind speeds $|u_h| (32+ \, m \, s^{-1})$ are referred to as TLVs to emphasize the fact that resolving the actual tornadoes likely will require even higher resolutions than currently used. All three grids used 53 vertical levels, with the spacing increasing from 20 m near the ground to 800 m at the model top located at 20 km. This vertical grid structure yields 7 (11) scalar$^2$ levels in the lowest 1 (2) km AGL with approximate grid spacings at 1 km and 2 km AGL of 220 m and 310 m, respectively. The first scalar model level is thus located at approximately 10 m AGL, and will hereafter be referred to as the “surface” when referencing scalar variables.

For a given grid, experiments are differentiated by the MP schemes/configurations employed. As in D10 we will use the following naming convention; the experiment names will follow the template $[dx][scheme]$, where $[dx]$ is the horizontal grid spacing with units and $[scheme]$ is the abbreviated MP scheme/configuration in capitals as listed in Table 1. On the outer 3-km grid, a single experiment (3kmMY3) is performed that uses the most sophisticated MY3 scheme. On the 1-km and 250-m grids, several experiments using different MP

$^2$The ARPS utilizes a standard Arakawa C-grid (Arakawa and Lamb 1977) where the scalar state variables are defined at the centers of grid boxes and the three velocity components on the faces.
schemes/configurations are performed (Table 1). Fig. 2 shows a schematic of the experiment design. The inner 1-km and 250-m grid experiments are stratified by the number of moments predicted in the MP scheme. Similar to the LINA and LINB experiments in D10 that were based on the Lin et al. (1983) 1M scheme, we perform two separate 1M experiments, denoted MY1A and MY1B, based on the 1M-MY scheme, on both 1-km and 250-m grids (each 250-m experiment is nested within the corresponding 1-km experiment). The MY1A experiments use the standard Marshall-Palmer (1948) exponential DSD for rain, with rain intercept parameter \( N_{0r} \) set to \( 8 \times 10^6 \) m\(^{-4} \), while MY1B reduces \( N_{0r} \) to \( 4 \times 10^5 \) m\(^{-4} \). This value, as discussed in D10, reduces the overall strength of the cold pool by shifting the rain DSD toward larger drops and correspondingly smaller evaporation rates, producing overall results similar to that seen in the MY2 and MY3 experiments in that study.

b. Microphysics scheme

As previously stated, all experiments use the MY-BMP scheme, which in its full implementation (selectable at runtime), predicts up to three moments of the assumed gamma size distribution for each of the hydrometeor categories of rain, ice crystals, snow, graupel, and hail (abbreviated r,i,s,g, and h, respectively), and up to two moments for cloud droplets. The gamma size distribution is given by

\[
N_\chi(D) = N_{0\chi}D^{\alpha\chi}\exp(-\lambda_\chi D),
\]

where \( N_\chi(D) \) is the number density of hydrometeors as a function of diameter \( D \), and \( N_{0\chi}, \alpha_\chi, \) and \( \lambda_\chi \) are the intercept, shape, and slope parameters, respectively, and the subscript \( \chi \) refers to any of the aforementioned hydrometeor categories. The moments predicted are the total number concentration \( N_{T\chi} \), the mass mixing ratio \( q_\chi \), and the radar reflectivity factor \( Z_\chi \),
proportional to the $0^{th}$, $3^{rd}$, and $6^{th}$ moments, respectively. Further details of this scheme can be found in MY05a,b, MY06a,b and D10. The shape parameter $\alpha_x$ is set to 0 for all categories in the 1M and 2M configurations (reducing the distributions to exponential). In the full 3M version of the scheme, all three free parameters of the gamma size distribution are allowed to vary independently. Since all three are seen to vary widely in observed rain DSDs (Ulbrich 1983), the 3M scheme, among bulk schemes, is the most flexible in this regard and potentially capable of representing a much wider range of DSDs than its 1M and 2M counterparts.

c. Data assimilation and forecast cycles

We performed hourly assimilation cycles from 1800 UTC to 0300 UTC on the 3-km grid first. This covers a period starting from approximately 2 h prior to the initiation of convection in Oklahoma to approximately 2 h after the major F5 (Fujita 1971) tornado swept through the Oklahoma City (OKC) area. The initial background field at 1800 UTC was taken from the 32-km North American Regional Reanalysis (NARR, Mesinger et al. 2006), and the boundary conditions for the 3-km grid are from the NARR at 3-hourly intervals. When available, the following conventional data were assimilated at each hour: upper-air soundings, wind profiles from the National Profiler Demonstration Network, Surface Aviation Observations, and Oklahoma Mesonet observations. In addition, visible and infrared satellite images from GOES 8 were assimilated through the ARPS cloud analysis system (e.g., Zhang et al. 1998) in order to build up the extensive cirrus canopy that was present over much of the southern Plains during the event. This cirrus canopy was found to be important in suppressing the development of early widespread convection in a previous modeling study of this event (Roebber et al. 2002), allowing for the development of relatively discrete and intense supercells later in the afternoon when breaks in the cloud cover moved over southwest Oklahoma. Speheger et al. (2002)
provided a detailed overview of the ensuing tornado outbreak, and the reader is referred to that study for further details. We adopt the lettering and numbering convention of Speheger et al. (2002) for the individual supercells and tornadoes in this event. For example, the long-track F5 tornado produced by the first supercell of the outbreak (storm A) was the ninth tornado produced by this storm and is labeled A9. Common errors between the simulations on the 1-km and 250-m grids, such as a high dewpoint bias in the warm sector (~ 3 K near the storms; not shown) likely arise mainly from the 3-km solution. However, the purpose of the 3-km grid was to provide a reasonable mesoscale environment, and thus serves mainly to provide boundary conditions for the one-way nested 1-km grid. An analysis of the mesoscale errors arising from the 3 km grid is outside the scope of this study. For these reasons, the results of the 3-km grid will not be further discussed.

On the 1-km grid, we performed 10-min assimilation cycles from 2100 UTC to 2250 UTC. This time period covers the entire developing phase of storms A and B as well as the early tornadic phase of storm A. The long-track F5 tornado A9 developed at approximately 2326 UTC, and thus assimilation cycles end approximately 36 min prior to its genesis. The frequent assimilation cycles on the 1-km grid aim to “build up” storms A and B within the mesoscale environment initially established by the hourly analyses on the 3-km grid and further improved by data assimilation on the 1-km grid. In addition to all the data used by the 3-km grid, reflectivity and radial velocity data from the Twin Lakes, Oklahoma WSR-88D radar (KTLX) were assimilated using the same 3DVAR and cloud analysis procedures on the 1-km grid. Due to the relatively coarse temporal frequency of assimilation we simply used the closest radar volume scan in time at the regularly spaced 10-min analysis times. In general, there was no more than a 2-min difference between the start of a given volume scan and the corresponding
analysis time. Experiments in which the temporal frequency of assimilation was varied between 5 and 15 minutes were also performed (not shown), and results were qualitatively similar to those presented here.

The complex cloud analysis procedure used is very similar to that reported in Hu et al. (2006a). Specifically, for all regions of observed reflectivity (after having been remapped to the ARPS grid) greater than 40 dBZ, an adjustment to the model thermodynamic profile was performed such that it represents a moist adiabatic profile diluted by mixing. This adjustment introduces thermal buoyancy and moistening in regions of reflectivity, encouraging updraft growth in a subsequent model forecast. In addition, hydrometeor fields were derived from the reflectivity field using the reflectivity formulations mostly after Smith et al. (1975). The cloud analysis package was originally developed for the ARPS 1M Lin scheme (Lin et al. 1983; Tao and Simpson 1993) and thus does not provide additional moments beyond the mixing ratios which are needed for the multi-moment schemes used in this study. As an initial implementation, we chose simply to diagnose the additional 0th and 6th moments using constant values of $N_0$ and $\alpha$, consistent with those of the Lin et al. (1983) scheme to ensure that the cloud analysis does not result in inconsistencies between the various predicted moments. A more robust method for handling multi-moment schemes in the complex cloud analysis procedure will need to be developed in the future. In practice, we found that the adjustment to the hydrometeor fields had much less of an impact on the subsequent forecast than the temperature and moisture adjustments, as 1) the model quickly adjusted to the imposed heating and moistening, and 2) much of the added hydrometeors tend to fall out to the ground quickly as precipitation (not shown).

The radial velocity data were first remapped to the ARPS grid using a pre-processing program which also included several automated quality-control procedures to de-alias folded
velocities, remove ground clutter, and despeckle noisy data (Brewster et al. 2005). The remapped data were visually inspected (no manual correction was necessary) and subsequently assimilated via the ARPS 3DVAR analysis procedure that includes a 2D mass divergence weak constraint (Hu et al. 2006b). The mass divergence constraint helps improve the analysis of the cross-beam component of the wind from the radial velocity observations by coupling the wind components together (Hu et al. 2006b). Additional experiments (not shown) in which radial velocity data were withheld from the assimilation cycles resulted in inferior forecasts of the storm tracks. Thus, as in Hu et al. (2006b), we found in this study that the best assimilation and subsequent forecast of the storms were obtained when both reflectivity and radial velocity were assimilated.

Finally, for each 1-km experiment, a forecast was launched from 2250 UTC to 0100 UTC. This forecast covered the time period from approximately 36 min prior to the genesis of tornado A9 to approximately 12 min after its dissipation. For the 250-m grid, a forecast was run out to 0100 UTC from the 15 min forecast (valid at 2305 UTC) of each 1-km experiment interpolated to the 250-m grid (Fig. 2).

3. Forecasts on the 1-km grid

a. Mesocyclone tracks and cold pool evolution

To quantitatively evaluate the forecast tracks, we performed 3DVAR analyses at 30-min intervals on the same 1-km grid including radar radial velocity data and use the analyses as “truth” for verifying the forecast mesocyclone tracks (the observed reflectivity over the forecast verification period was remapped and smoothed for plotting purposes only). The locations of the mid-level (~3 km AGL) mesocyclone centers (defined as the scalar grid point with maximum
vertical vorticity $\zeta$ at 30-min intervals starting at 2300 UTC in the analyses and forecasts are plotted in Fig. 3, together with the observed tornado damage track (obtained from D. Speheger and S. Rae 2009 via personal communication), and the 30 dBZ reflectivity contours.

Substantial differences in the forecast tracks exist across the experiments, due mostly to differences in the translational speed of the mesocyclone. In general, as is also seen from the forecast mesocyclone position errors (Fig. 3f), the experiments using multi-moment MP outperform the 1M experiments in terms of the mesocyclone track. Between the two 1M experiments, a better forecast is produced by 1kmMY1B due to the reduced value of $N_\theta$ (see Table 1), skewing the DSD toward larger drops and reducing the evaporation rate of rainwater. However, all experiments exhibit a consistent eastward displacement of the forecast mesocyclone from the observed that may be related to other sources of error, particularly those in the initial conditions.

In each 1-km experiment, the qualitative structure of storm A is established quite well in the early forecast period, with a classic (simulated) radar presentation at 2330 UTC (40 min forecast). (Fig. 4). However, the reflectivity structure at this time differs substantially between the different experiments and from the remapped observed reflectivity (Fig. 4e). The same is true of the cold pool structure$^3$ (Fig. 5), with 1kmMY1A exhibiting the largest and strongest

$^3$ We use equivalent potential temperature $\theta_e$ to visualize and qualify the storm cold pools as in D10, despite the fact that this variable is not a proxy for buoyancy. $\theta_e$ is nearly conserved for pseudoadiabatic motions and is thus a useful proxy for the upper limit of vertical parcel displacements in environments where it decreases with height; relatively lower values at a given
cold pool (Fig. 5a). The 2M and 3M experiments (1kmMY2 and 1kmMY3; Fig. 4c,d, respectively) both exhibit larger forward flanks than either of the two 1M experiments (Fig. 4a,b), or the observed storm (Fig. 4e), and have very weak cold pools (Fig. 5c,d). Similar behavior was noted with the MY scheme in the idealized simulations of D10 and Wainwright et al. (2014). These studies attributed this difference in behavior at least partially to the combination of the action of size sorting in the multi-moment scheme and the assumed relatively low fall speed curve for graupel. These effects both allow greater amounts of low-density small graupel to advect further downwind (eastward) from the storm updraft, broadening the forward flank region. Both the reflectivity and cold pool structure in 1kmMY1B is intermediate between 1kmMY1A and the two multi-moment experiments (Fig. 4b,5b, respectively).

At later times (as typified by the 100 min forecast time; 0030 UTC), the reflectivity structure of the storms in 1kmMY1A and to a lesser extent 1kmMY1B (Fig. 6a,b) has departed considerably from the observed structure (Fig. 6e), with the orientation of the forward flank changing from an E-W configuration to more of a SSW-NNE configuration. In contrast, the structure in 1kmMY2 and 1kmMY3 (Fig. 6c,d) more closely resembles the observations (Fig. 6e) in both shape and orientation, although the overall size is still exaggerated. The change in structure of the 1M experiments is associated with the continued expansion and intensification of the cold pool (Fig. 7a,b), whereby the stronger outflow has pushed out further east in a storm-relative sense. In contrast, the cold pools in the multi-moment experiments remain weak (Fig. 7c,d).

level signify that at least some descent from altitudes corresponding to these values has taken place. We refer the reader to Markowski et al. (2002) for further discussion.
Thus, there is a large effect on the cold pool of either reducing the fixed $N_0$, in the 1M scheme, or predicting additional moments (which effectively also reduces $N_0$; see section 3c below). The composite minimum surface equivalent potential temperature $\theta_e$ [computed using the formula of Bolton (1980)] “swath” shown in Fig. 8\textsuperscript{4} gives an indication of the overall spatiotemporal evolution of the cold pool in each experiment. In 1kmMY1A (Fig. 8a), the strong cold pool appears to cause the storm to move faster than the observed one and those in the other experiments, particularly toward the end of the forecast period (as also seen in the large positive slope of the position error: black line in Fig. 3f). MY06b found similar behavior with their real-data sensitivity study of a supercell. Bunkers and Zeitler (2000) and Zeitler and Bunkers (2005) note that this mechanism for supercell propagation is not quantitatively well constrained. However, the eastward-directed difference in the track of the storm in 1kmMY1A relative to the other experiments is consistent with this mechanism of propagation. That is, the propagation of the gust front out ahead of the storm updraft continually forces new convective development on the southeast flank of the supercell, resulting in a component of motion of the supercell toward the east or southeast. This effect is in addition to the more dominant components of motion associated with advection by the mean wind and the dynamic interactions of the storm updraft and mesocyclone with the ambient shear (Zeitler and Bunkers 2005). To visualize this structural difference, we computed a storm-relative composite over the period from 2330-0100 UTC for each experiment. We calculate the composite by tracking the low-level ($\sim 186$ m AGL) updraft centers at 5 min intervals during this period (not shown). The updraft centers are computed as the location of maximum vertical velocity at the given height, taking

\textsuperscript{4} Plots in this and subsequent similar figures are produced by taking the maximum or minimum at each grid point of the field in question over the duration in question.
care to exclude maxima not associated with storm A. We then temporally average the model fields in a 24x24 km² region centered on the (translating) low-level updraft. Examination of individual times (not shown) reveals that the composites are good representations of the storm structure across most of the period, although as described above there is a general trend toward a stronger cold pool with time in 1kmMY1A. We present contour plots of the composite low-level (~186 m AGL) and mid-level (~3100 m AGL) updrafts, along with composite surface $\theta_e$ and wind vectors for each experiment in Fig. 9. Relative to the other experiments, the updraft in 1kmMY1A (Fig. 9a) has a larger southeast-to-northwest tilt with height and a gust front that tends to propagate further east of the main updraft. Interestingly, a hint of a secondary gust front closer to the main updraft is present in the composites of 1kmMY1A (and to a lesser extent in 1kmMY1B, Fig. 9b) but not the other experiments (though they may be present at individual times). Similar structures have also been noted and described in several recent observational studies of supercells (Wurman et al. 2007; Marquis et al. 2008; Wurman et al. 2010; Kosiba et al. 2012; Lee et al. 2012; Markowski et al. 2012; Marquis et al. 2013; Skinner et al. 2014).

b. Comparison with Oklahoma mesonet observations

To evaluate further the cold pool in the 1-km experiments, we make a comparison with Oklahoma Mesonet (Brock et al. 1995) observations. Because the Mesonet mean station spacing is about 30 km, the network’s spatial resolution is still too coarse to determine the exact location and gradients across the outflow boundary. Similar to Schenkman et al. (2011), we choose to examine the time series of thermodynamic variables at a given Mesonet station, which gives us an idea of the evolution of such variables as the storm passes over the station. The Spencer, Oklahoma station was located in the path of storm A’s precipitation core during the forecast period and thus experienced the outflow of the storm (Fig. 8, see also Fig. 7 in D10). Therefore,
we chose to single this station out for investigation. Shown in Fig. 10 are the time series of observed surface temperature $T$, dewpoint temperature $T_d$, $\theta_e$, and accumulated precipitation. Also shown are the corresponding time series extracted from the 1-km forecasts. The observed $T$ and $T_d$ series in Fig. 10a show relatively small changes after the onset of precipitation; a slight drop in $T$ and corresponding increase in $T_d$ is consistent with evaporative cooling by rain (note that the onset of precipitation in the 1km experiments precedes that of the observations by ~30 min, c.f. position increments in Fig. 3). The observed $\theta_e$ (solid black line in Fig. 10b) remains relatively constant during this time, consistent with small parcel displacements in the vertical and/or little or no entrainment of surrounding environmental air during processing by the storm updraft and downdraft (Markowski et al. 2002).

In contrast, the corresponding model time series from different experiments differ significantly from each other and from the observations. In particular, 1kmMY1A and 1kmMY1B show significant decreases in $T$, $T_d$, and $\theta_e$ coincident with the onset of precipitation (compare blue and green lines in Fig. 10), with 1kmMY1A showing the largest $\theta_e$ perturbation of ~ -19 K (blue line in Fig. 10b). In contrast, both 1kmMY2 and 1kmMY3 show relatively constant $T$, $T_d$, and $\theta_e$ during and after the onset of precipitation, more similar to the observations. For the accumulated precipitation (Fig. 10c), 1kmMY1A and 1kmMY1B have significantly higher precipitation totals than the observations (by nearly a factor of 2) by the end of the period shown, while 1kmMY2 and 1kmMY3 display very similar total accumulated precipitation. No attempt was made to correct the time series for differences in storm propagation between the simulations and observations. As such the earlier onset of cooler temperatures and substantial precipitation in 1kmMY1A and 1kmMY1B (blue and green lines in
relative to the observations (black lines in Fig. 10) is concomitant with the faster storm motion and further northeastward location relative to the observed storm (Fig. 3). The same is true but to a much lesser extent for 1kmMY2 and 1kmMY3. Nevertheless, the core of each simulated storm traversed the Spencer site in a qualitatively similar manner as the observed storm. Thus, differences in thermodynamic and precipitation characteristics seen in Fig. 10 are largely a function of intrinsic differences in storm microphysics rather than differences in location relative to the Spencer site. In any case, a much better agreement with the observations is obtained in the predicted surface thermodynamic and precipitation characteristics by the multi-moment versions of the MY scheme.

c. Microphysical and downdraft evolution

The large differences in cold pool strength and structure between the experiments have their root cause in differences in the latent cooling from hydrometeor phase changes in the downdrafts. To analyze these effects, we first define a moving 48x48 km$^2$ subdomain that tracks the mid-level (~2925 m AGL) mesocyclone center of storm A at 5 min intervals from the period 2330-0100 UTC for each experiment. We consider all grid points within this subdomain with vertical velocity $w < -0.5$ m s$^{-1}$ and height AGL $z < 4$ km (near and below the melting level) for each of the subsequent analyses.

Fig. 11 shows time-height plots of horizontal minimum (within this storm-following subregion) $\theta_e$ (top subpanels) and $w$ (bottom subpanels) for each experiment. Experiments 1kmMY1A and 1kmMY1B produce overall stronger downdrafts that episodically penetrate to lower levels than 1kmMY2 and 1kmMY3 (compare bottom panels of Fig. 11a,b with Fig. 11c,d). The stronger downdrafts in 1kmMY1A and 1kmMY1B are associated with overall lower values of minimum $\theta_e$, as would be expected from deeper descent from the mid-troposphere (Fig.
11a,b, top subpanels; see also Fig. 3 in D10). The minimum $\theta_e$ in the lowest ~500 m AGL in
1kmMY2 and 1kmMY3 (Fig. 11c,d top plots) is also on the order of 10-20 K greater than in
1kmMY1A and 1kmMY1B. The weaker, more elevated nature of the downdrafts in 1kmMY2
and 1kmMY3 are consistent with the Oklahoma Mesonet observations described previously, in
that the surface air in the precipitating regions is consistent with evaporatively-cooled boundary
layer air that did not have its origin from higher altitudes.

We show the total amount of latent cooling within the moving subdomain from different
hydrometeor phase change source terms summed over the budget period 2330-0100 UTC for
each of the 1-km experiments in Fig. 12. The dominant contributions to latent cooling in the
downdrafts come from cloud and rain evaporation and hail melting; all other latent cooling
processes, including ice, snow, graupel, and hail sublimation, and ice, snow, and graupel melting,
are aggregated in the red bars in Fig. 12. As can be seen, 1kmMY1A has substantially more
latent cooling from all processes than in the other experiments, particularly that from rain
evaporation (where it is greater by approximately a factor of 4). Simply lowering the fixed $N_{0r}$
by a factor of 20 in 1kmMY1B produces a latent cooling budget very similar to those of
1kmMY2 and 1kmMY3. These results are consistent with the idealized simulations of D10 (see
their Fig. 8).

To investigate further, we plot in Figs. 13-16 vertical profiles and time series of quantities
related to cloud, rain, and hail for each experiment (valid for the aforementioned downdraft
subdomain). Again, experiment 1kmMY1A (blue curves) has by far the most amount of cooling,
the majority of which comes from the increased evaporation of rain (Fig. 13b). This result is
commensurate with the largest magnitudes of $N_{0r}$ (Fig. 14d) and horizontally-averaged $q_r$ (Fig.
14b), and the largest total mass of rain (Fig. 16b) in this experiment as compared to the others,
and these differences persist throughout the forecast period (Fig. 15b,16b). Both multi-moment experiments exhibit profiles of average $N_{0r}$ (red and purple curves in Fig. 14d) that are intermediate between those of the fixed values of the 1M experiments, but are closer to 1kmMY1B (cyan curves). These results are consistent with Wainwright et al. (2014) who examined the ability of an appropriately “tuned” 1M scheme (with either fixed or diagnostic- $N_{0r}$ ) to reproduce certain features of a 2M scheme.

Turning to cloud water evaporative cooling (Fig. 13a), the experiments show somewhat less variation, but 1kmMY1A still has the most for all heights and for most of the budget period (Fig. 15a), and exhibits slightly more cloud mass (Fig. 14a, Fig. 16a). Experiments 1kmMY1A and 1kmMY1B differ only by the different assumed fixed $N_{0r}$, and thus the differences here should be due to differences in accretion rates of cloud by rain and in nonlinear interactions with the storm dynamics. For example, the stronger downdrafts in 1kmMY1A (Fig. 11) tend to induce more entrainment and evaporation of cloud water in the downdraft region.

The vertical profiles of latent cooling by hail melting (Fig. 13c) differ substantially from each other depending on whether the experiment used 1M or multi-moment MP. Both 1kmMY1A and 1kmMY1B have comparable magnitudes of total melting by hail which is consistently approximately two times greater than 1kmMY2 and 1kmMY3 (Fig. 15c). The multi-moment experiments show a marked increase of average $N_{0h}$ with height (Fig. 14e) that

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5 For 1kmMY3, as in D10, we compute the normalized intercept parameter $N_0^*$ (Testud et al. 2001), which is the $N_0$ of the corresponding exponential distribution with the same mass-weighted mean diameter and hydrometeor mass content.
quickly becomes substantially larger than the assumed fixed $N_{0h} (4.0 \times 10^4 \text{ m}^{-4})$ for the 1M experiments. The greater magnitude of melting on the part of the 1M experiments can be explained as a combination of an overall greater mass of hail (Fig. 16c), and the smaller $N_{0h}$ for hail (Fig. 14e). On the one hand, all else being equal, smaller $N_{0h}$ would decrease the melting rates owing to the corresponding decrease in total surface area of the distribution. On the other hand the smaller average $N_{0h}$ in the 1M experiments is associated with higher terminal velocities. As such, the hail is able to fall further into the much warmer lower levels before completely melting (Fig. 14c). Indeed, both 1kmMY2 and 1kmMY3 have a peak in melting at ~2.25 km AGL with much smaller magnitudes below, whereas both 1kmMY1A and 1kmMY1B have a broader peak centered near 1.5 km AGL (Fig. 13c). These results are broadly consistent with those of Gilmore et al. (2004), who performed a set of idealized supercell simulations in which they varied the magnitude of $N_{0h}$. In addition to a similar downward shift in the vertical profile of cooling as $N_{0h}$ was decreased, they found that the higher the terminal fall speeds, the less the residence time as the hail falls through the environmental wind profile. This in turn leads to less horizontal advection and a more spatially concentrated region of melting. Additionally, we note that the warmer environmental temperatures in the low-levels would directly contribute to increased local melting rates. Finally, the stronger downdrafts in 1kmMY1A and 1kmMY1B (Fig. 11) may contribute by (vertically) advecting more hail mass to lower levels.

In general, the results of the MP budget analysis in the downdrafts are consistent overall with the idealized simulations of D10. The larger magnitudes of evaporation in the 1M simulations of D10 were also found to be due to a combination of greater amounts of rain and smaller mean drop sizes (or larger $N_{0r}$). The lack of size sorting in the 1M scheme is a partial
explanation: smaller mean diameters (associated with enhanced evaporation potential) are effectively able to sediment to lower levels owing to the single mass-weighted terminal velocity used. On the other hand, the multi-moment schemes allow for a size-sorting mechanism via differential sedimentation of the predicted moments whereby the mean volume diameter $D_{mx}$ increases toward the ground (see MY05a; D10; Milbrandt and McTaggart-Cowan 2010; and Kumjian and Ryzhkov 2012 for further discussion), resulting in overall less evaporative cooling in the low levels. Additionally, the flexibility that comes from predicting at least two moments of the rain and hail distributions allows for the initiation of rain distributions characterized by relatively large drops from melting of hail; in a 1M scheme, the rain and hail sizes are both completely determined by a single predicted moment (typically the mixing ratio) along with the assumed fixed (or possibly diagnosed) $N_0x$ (see discussion in Wainwright et al. 2014).

4. Results of 250-m grid simulations

We performed similar microphysical analyses for the 250-m experiments as for the 1-km ones. The 250-m results are consistent with those of the 1-km grid; for brevity, we focus in this section on the impact of the MP scheme on the prediction of TLV tracks as compared to the observed tornado track. Strong TLVs form within storm A in each of the 250-m experiments but with substantial differences in the behavior of the vortices, as seen in Figs. 17,19 for composite horizontal wind speed $|u_h|$ and minimum $\theta_e$ over the period 2320-0100 UTC (900-6900 s model time). To delineate individual TLVs, we impose the following criteria. A TLV is assumed to be present at a given time if the maximum surface wind speed $|u_h|_{max}$ within a 4 km radius of the location of maximum surface vorticity $\zeta_{max}$ is at least 32 m s$^{-1}$ for a period of at least 2 min. A gap of up to 2 min is allowed in which $|u_h|_{max}$ can dip below this threshold. While these criteria
have limitations, particularly 1) not differentiating between strong outflow winds and winds associated with the TLV itself, and 2) inability to discriminate between multiple simultaneous TLVs, we confirmed through inspection (not shown) that these limitations were not a factor in our particular simulations. Also, while other criteria for determining the presence of TLVs exist, such as those based on the Okubo-Weiss number (e.g., Markowski et al. 2011), these are not without their own limitations (such as difficulty in mapping wind speed thresholds to particular magnitudes of the given metric). Our chosen wind speed threshold corresponds to the lower boundary of F1 on the original Fujita scale (Fujita 1971). It is also near the lower end of the Enhanced Fujita (EF) scale (WSEC 2006), and is very similar to that used by Schenkman et al. (2012). The number, duration, cumulative track length, and strength of TLVs in each simulation using these criteria are tabulated in Table 2, along with corresponding data for the observed F5 tornado [A9 in Speheger et al. (2002)]. The time series of $\zeta_{\text{max}}$ and $|\mathbf{u}|_{\text{max}}$ for each experiment are shown in Fig. 18 and the portions of each curve corresponding to a TLV detection are highlighted in bold.

The 250-m experiments reflect the trend of the 1-km experiments in regard to cold pool strength, although there appears to be slightly less overall sensitivity in regards to the minimum $\theta_e$ perturbations and overall cold pool size across the experiments (compare Fig. 8 and Fig. 19). Future work may address this possible resolution-dependence on microphysics and cold pool dynamics. It can be seen that the farthest southeast position of the gust front represented by the sharp gradient in minimum $\theta_e$ (Fig. 19) corresponds well with the forecast TLV track, represented by the 32 m s$^{-1}$ composite $|\mathbf{u}|$ contour in black in each panel. In 250mMY1A, the behavior of the forecast TLV is qualitatively different from those in all the other experiments; four relatively weak, short-lived TLVs are produced in the experiment (Table 2), although the
earliest vortex swath is quite close to the beginning part of the actual tornado track (Fig. 17a).

This experiment displays the weakest peak $\zeta_{\text{max}}$ and $|u_h|_{\text{max}}$ magnitudes (~0.2-0.25 s$^{-1}$, and ~49 m s$^{-1}$, respectively, blue lines in Fig. 18).

In contrast, both 250mMY1B and 250mMY3 produce long-track TLVs with track lengths qualitatively similar to that of tornado A9 (Fig. 17b,d, respectively). The single detected TLV in 250mMY3 had a peak $|u_h|_{\text{max}}$ of 77 m s$^{-1}$ (EF4) closest to the observed intensity of tornado A9 (F5), with peak $\zeta_{\text{max}}$ near 0.4 s$^{-1}$ (purple lines in Fig. 18). The duration of the TLV in this experiment was 3930 s (or 65.5 min, Table 2), which is similar to the observed tornado’s duration (~90 min, Speheger et al. 2002). It is worth noting at this point that the TLV in 250mMY3 was still in progress and rather intense at the cessation of the forecast, and likely would have continued for some time longer had the forecast been extended. On the other hand, the TLV-genesis in 250mMY3 was delayed by approximately 30 min when compared to the observed tornado (genesis at 2323 UTC, first vertical black line in Fig. 15). The TLVs in 250mMY1B had a peak surface wind of 56 m s$^{-1}$ (EF2, Fig. 18, cyan line, Table 2). This experiment also exhibited a relatively long duration of $\zeta_{\text{max}} > 0.1$ s$^{-1}$ and $|u_h|_{\text{max}} > 32$ m s$^{-1}$ which were nevertheless ~ 30% less than that seen in 250mMY3. However, it shows the best overall agreement with the observed temporal window of tornado A9 (vertical black lines in Fig. 18).

Indeed, the gap between the two later TLVs in 250mMY1B (see Table 2) just barely evades the aforementioned TLV criteria: if they were to be relaxed slightly, a single TLV with a duration of ~57 min and cumulative track length of ~45 km would result. Finally, experiment 250mMY2 (Fig. 17c) produces two TLVs with intensities comparable to the single intense TLV in 250mMY3 [peak surface wind of 70 m s$^{-1}$ (EF3), red lines in Fig. 18, Table 2], but shows cyclic behavior unlike what was observed, and also exhibited delayed genesis similar to that of
250mMY3. One can see, in any case, a trend toward more intense and/or longer track vortices when moving from the 1M version of the MY scheme to the 2M and 3M versions.

Turning to the minimum $\theta_e$ composites (Fig. 19), large differences in cold pool strength and area are seen that are qualitatively similar to the 1-km experiments (Fig. 8) but exhibit more details consistent with the higher spatial resolution. Experiments 250mMY1A and 250mMY1B (Fig. 19a,b, respectively) show a nearly continuous swath of lower $\theta_e$ (< 345 K) north of the gust front position (given approximately by the green 342 K contour in Fig. 19), while 250mMY2 and 250mMY3 display a relative maximum in $\theta_e$ between lower values to the north and slightly lower values to the south, closer to the position of the simulated TLV track (given by the black 32 m s$^{-1}$ contour). This southern swath represents the RFD, and tends to be at its most intense when the TLV is intense. Inspection of individual times during the evolution of the cold pool in each experiment reveals a tendency for the regions of coolest outflow from the forward and rear flank downdrafts in 250mMY1A and 250mMY1B to merge over much of the duration of the forecast, while they remain more separated in 250mMY2 and 250mMY3. To illustrate this, we show plots of surface $\theta_e$, horizontal wind vectors and $\zeta$ at the time of peak intensity of one of the TLVs in each experiment in Fig. 20. In 250mMY1A, and to a lesser extent, 250mMY1B (Fig. 20a,b, respectively), there is a well-defined SSW-NNE-oriented boundary separating air with relatively high $\theta_e$ (345+ K) to the east from outflow air with $\theta_e$ on the order of 325-330 K to the west. This boundary appears to be an instance of the so-called left flank convergence boundary (LFCB) identified by Beck and Weiss (2012) in their high-resolution idealized supercell simulation study. In 250mMY1A in particular, the lower $\theta_e$ air to the west of the boundary has infiltrated the TLV region (Fig. 20a). In contrast, in both 250mMY2 and
260mMY3 (Fig. 20c,d, respectively), neither the LFCB, nor any strong east-west gradient in $\theta_e$, is apparent in this region. Instead, the lower $\theta_e$ values associated with the storm outflow appear in two separate regions: 1) much further west, to the northwest of the precipitation core, and 2) in a narrow zone just behind the rear-flank gust front (RFGF). In both cases, the TLV is embedded in a nearly uniform region of high $\theta_e$ air uniform more typical of the near-surface inflow (350-355 K). Moreover, the portions of the minimum-$\theta_e$ swath in 250mMY1A that correspond with the TLV track (black contours in Fig. 19a) display higher values than elsewhere along the swath, suggesting a positive correlation between higher-$\theta_e$ outflow air and the presence of a TLV for this experiment. However, this correlation is not apparent for the other experiments, possibly because the “baseline” outflow has $\theta_e$ values already similar to those of the inflow.

5. Summary and Conclusions

In this study, we investigated the sensitivity of the prediction of the 3 May 1999 Oklahoma City, Oklahoma tornadic supercell and its associated tornadoes to the use of 1M, 2M and 3M versions of the Milbrandt and Yau (2005b) BMP in a real-data context by assimilating both conventional and Doppler radar data. The data assimilation and numerical prediction setup utilized telescoping nested grids from a 3-km grid spacing for the outermost grid, to a 1-km grid on which radar data assimilation was performed, and to a 250-m grid that attempted to resolve tornado-like vortices. The data assimilation procedure used the intermittent 3DVAR and cloud analysis data assimilation strategy documented in Hu et al. (2006a; b); conventional observations were first assimilated on the 3-km outer grid to capture the mesoscale environment in which convective storms develop, and in addition radar radial velocity and reflectivity data were assimilated at 10-min intervals on the nested 1-km grid to initialize the developing supercell
storms. We then performed forecasts on the nested 250-m grid from interpolated, short 1-km grid forecasts. We examined the sensitivity of the predicted storm on the 1-km grid, and predicted development and evolution of the tornado-like vortices on the 250-m grid to versions of the BMP. To our knowledge this is the first study to investigate the sensitivity of tornado-scale simulations to single- vs. multi-moment microphysics in a real-data framework.

Systematic differences were found between the experiments that used different BMP. In general, the multi-moment simulations better captured the behavior of the storm, both from a mesocyclone track and tornadic activity perspective. The main conclusions of this study are summarized as follows:

1) On the 1-km grid, the track of the 3 May 1999 Oklahoma City, Oklahoma tornadic supercell was predicted better by the multi-moment versions of the MY scheme, with the 3M version (experiment 1kmMY3) performing the best and having the lowest overall position errors for the simulated mesocyclone. The 3M version also produced a relatively weak cold pool, which was in a much better agreement with the available surface observations. In contrast, the 1M version with the default rain intercept parameter (named 1kmMY1A) produced the worst overall track and an overly strong cold pool. The 1M experiment with the reduced rain intercept parameter (1kmMY1B) and the 2M experiment (1kmMY2) produced results in-between these two extremes. The accumulated precipitation at the Spencer, Oklahoma Mesonet site was also better predicted by the multi-moment schemes.

2) An analysis of the microphysical processes responsible for latent cooling in the downdraft regions of the storms in the 1-km experiments revealed that experiment 1kmMY1A produced evaporation of rain in the downdrafts integrated over the duration
of the forecast that was roughly a factor of 4-5 times greater than those of the 1kmMY1B, 1kmMY2, and 1kmMY3 experiments. Cooling from other processes displayed similar trends, but was of secondary importance. The increased evaporation of rain in 1kmMY1A was due both to the increased rain mass in the experiment as well as the larger magnitude of the intercept parameter used (see also the discussion in D10), which in turn led to overall stronger and deeper downdrafts and stronger cold pools relative to other experiments. The stronger cold pool in 1kmMY1A, and to a lesser extent, 1kmMY1B resulted in a greater east-to-west slope of the low level updraft and faster forward propagation consistent with the results of Milbrandt and Yau (2006) and Snook and Xue (2008).

3) On the 250-m grid, the multi-moment schemes performed better than the 1M version with the default rain intercept (250mMY1A) in producing longer-lived and stronger TLVs, although there was significant variability in the predicted track. Though again we point out that the 250-m grid is still too coarse for tornadoes to be resolved fully, nevertheless the maximum intensity of the TLV in the experiment using the multi-moment scheme (250mMY3) was comparable to the observed intensity of the tornado (EF4 vs. EF5\textsuperscript{6} on the Enhanced Fujita scale). Experiment 250mMY2 produced intense vortices (EF3), but displayed a cyclic behavior that differed from that observed. However, both multi-moment experiments exhibited substantially delayed genesis relative to the observed tornado, by approximately 30 min in both cases. Experiment 250mMY1B also produced

\textsuperscript{6} Technically the actual tornado was rated F5 on the original Fujita (1971) scale, but we are treating this as equivalent to an EF-5 rating.
a long-track TLV similar to the observed tornado, but was substantially weaker (EF2) than in 250mMY3. However, the timing of genesis agreed quite well with the observed tornado. Finally 250mMY1A produced the shortest-lived and weakest TLVs (EF1). Thus, in agreement with the idealized simulations of Snook and Xue (2008), changing the magnitude of the rain intercept parameter in the 1M schemes had a profound effect on the TLV behavior, with smaller (larger) values of $N_0r$ leading to stronger (weaker) TLVs.

4) The qualitative differences in cold pool strength on the 1-km grid were also reflected on the 250-m grid. Regardless of the cause for larger raindrops and/or hailstones in the low-levels in previous and the current study, the results in the context of supercell tornadogenesis appear to be similar: weaker cold pools are associated with more intense and longer-lived TLVs in the simulated supercells, which agrees well with available observations (e.g., Markowski 2002; Markowski et al. 2002; Shabbott and Markowski 2006; Lee et al. 2012; Markowski et al. 2012).

We wish to emphasize that the actual details of the simulated individual TLVs can be sensitive to other details of the forecast experiment setup. During the course of this work, several other simulations similar to the ones reported on in this study were performed, using somewhat different data assimilation strategies (i.e. different window lengths, initial times, and assimilation frequencies) and forecast initial times. In each case, even though the details of the simulated storms and TLVs differed, the same basic trend across BMPs as noted above was observed among the experiments (not shown). For these reasons, we believe, at least for this case, that the general impacts of the BMPs on the storm and TLV simulations are robust. Future work will investigate the sensitivity of TLV prediction to microphysics through an ensemble-based approach that will help address sensitivity to initial conditions.
The above findings broadly indicate that improving the BMP directly improves the simulation and prediction of supercell thunderstorms and their associated tornadoes, at least in this case for the particular choice of BMP. The BMP substantially modulates the storm cold pool strength, which apparently has substantial effects on the structure and evolution of the simulated supercell storms and the genesis and evolution of embedded tornadoes. In this case, we demonstrated an improvement (better agreement with observations) with increasing prognostic detail on the hydrometeor size distributions: both on the storm scale in regards to the storm track and cold pool properties, and on the (near) tornado scale in regards to the intensity and duration of the simulated TLVs. On the other hand, sensitivity to the choice of multi-moment BMP was not assessed in this study, but may be examined in the future. For example, recent studies (Morrison and Milbrandt 2011; Van Weverber 2011; Morrison et al. 2012) have shown that 2M schemes are very sensitive to the density and fall speed characteristics of the large rimed ice category (i.e., graupel or hail), as well as the rain drop breakup parameterization. Differences in the treatment of these and other processes can lead to differences in storm structure and cold pool characteristics that are qualitatively as large as the differences between the 1M, 2M, and 3M experiments in our study.

The relationship between the cold pool intensity and tornadogenesis potential has been a topic of several past studies (Markowski 2002; Markowski et al. 2002; Markowski et al. 2003; Lerach et al. 2008; Snook and Xue 2008; Lee et al. 2012; Lerach and Cotton 2012) and continues to be a topic of active research. The broad consensus of these studies is that tornadogenesis and maintenance are favored when the cold pool is relatively weak [i.e. temperatures differing by O(1-10 K) from inflow values] and becomes increasingly suppressed as outflow temperatures become colder. However, the exact processes by which the cold pool...
affects tornadogenesis and behavior, and in particular how the coldness or buoyancy of the low-level parcels that feed the low-level vortex affects the intensification of the vortex are still unclear. It has been suggested (Markowski et al. 2008) that there is an optimal strength of the cold pool that provides the most favorable conditions for tornadogenesis. On the one hand baroclinic vorticity generation within the cold pool can provide a near-ground vorticity source, but on the other the cold pool shouldn’t be so strong as to overwhelm vertical stretching through negative buoyancy and possibly also disconnecting the incipient tornado from the support of the low-level updraft and mesocyclone (Snook and Xue 2008).

In part II of this study (Dawson et al. 2014), we will address some of the above questions by examining the low-level pressure gradient force and the buoyancy field (which is directly linked to the coldness and hydrometeor loading of the cold pool air) in the near-TLV environment within the experiments using different BMPs, and try to determine quantitatively how the low-level buoyancy field affects the formation and intensification of the low-level TLVs in two of the 250-m experiments (250mMY1A and 250mMY3). This is accomplished by calculating the dynamic pressure gradient and buoyancy forces along parcel trajectories, and by studying their relationships with the behavior of the simulated TLVs.

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Table 1. List of microphysics schemes and their configurations used in the 1-km and 250-m experiments.

<table>
<thead>
<tr>
<th>Microphysics scheme/configuration</th>
<th>Description</th>
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<tr>
<td>MY1A</td>
<td>Single-moment MY scheme with rainwater intercept parameter $N_{0r} = 8 \times 10^6 \text{ m}^{-4}$</td>
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<tr>
<td>MY1B</td>
<td>Single-moment MY scheme with rainwater intercept parameter $N_{0r} = 4 \times 10^5 \text{ m}^{-4}$</td>
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<tr>
<td>MY2</td>
<td>Double-moment MY scheme (mixing ratios $q_x$ and total number concentrations $N_{fx}$ predicted)</td>
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<tr>
<td>MY3</td>
<td>Triple-moment MY scheme ($q_x$, $N_{fx}$ and reflectivity factor $Z$ predicted)</td>
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Table 2. List of detected TLVs including their start and end times, duration, cumulative track length and strength, for each of the 250-m experiments. Also listed are corresponding data for the observed tornado A9 (taken from Speheger et al. 2002). Note that the 3 May 1999 tornado was rated F5 under the original Fujita scale but this can be considered equivalent to an EF5 rating under the new scale for the purposes of this study.

| Experiment   | Number of TLVs | Start-End/Duration/Track length (UTC/min/km) | $\left| u_\text{max} \right|$ (m s$^{-1}$)/EF# |
|--------------|----------------|---------------------------------------------|-------------------------------------|
| 250mMY1A     | 4              | 2323-2332/9.0/8.0                            | 49.1/EF1                           |
|              |                | 2336-2342/6.0/5.3                            | 43.6/EF1                           |
|              |                | 0028-0035/6.5/3.7                            | 42.5/EF1                           |
|              |                | 0046-0050/4.5/3.8                            | 42.6/EF1                           |
| 250mMY1B     | 3              | 2326-2337/11.0/8.8                           | 45.2/EF1                           |
|              |                | 2341-0009/27.5/22.8                          | 48.9/EF1                           |
|              |                | 0011-0041/29.5/22.3                          | 55.7/EF2                           |
| 250mMY2      | 2              | 2353-0011/17.5/17.2                          | 70.3/EF3                           |
|              |                | 0021-0100/38.5/24.1                          | 69.7/EF3                           |
| 250mMY3      | 1              | 2354-0100/65.5/44.3                          | 76.7/EF4                           |
| Observed (tornado A9) | 1          | 2326-0048/82.0/59.5                          | NA/(E)F5                           |
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