Effects of Under-Resolved Convective Dynamics on the Evolution of a Squall Line

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ABSTRACT

Simulations of a squall line observed on 20 May 2011 during the Midlatitude Continental Convective Clouds Experiment (MC3E) using 750- and 250-m horizontal grid spacing are performed. The higher-resolution simulation has less upshear-tilted deep convection and a more elevated rear inflow jet than the coarser-resolution simulation in better agreement with radar observations. A stronger cold pool eventually develops in the 250-m run; however, the more elevated rear inflow counteracts the cold pool circulation to produce more upright convective cores relative to the 750-m run. The differing structure in the 750-m run produces excessive midlevel front-to-rear detrainment, reinforcing excessive latent cooling and rear inflow descent at the rear of the stratiform region in a positive feedback. The contrasting mesoscale circulations are connected to early stage deep convective draft differences in the two simulations. Convective downdraft condensate mass, latent cooling, and downward motion all increase with downdraft area similarly in both simulations. However, the 750-m run has a relatively greater number of wide and fewer narrow downdrafts than the 250-m run averaged to the same 750-m grid, a consequence of downdrafts being under-resolved in the 750-m run. Under-resolved downdrafts in the 750-m run are associated with under-resolved updrafts and transport mid–upper-level zonal momentum downward to low levels too efficiently in the early stage deep convection. These results imply that under-resolved convective drafts in simulations may vertically transport air too efficiently and too far vertically, potentially biasing buoyancy and momentum distributions that impact mesoscale convective system evolution.

1. Introduction

Squall lines are a common mesoscale convective morphological mode that have been frequently studied because of their propensity to organize and propagate over a variety of scales up to 1000 km or more across tropical and midlatitude environments, causing significant hydrological and severe weather impacts (e.g., Johns and Hirt 1987; Doswell et al. 1996; Trapp et al. 2005). They also commonly produce distinguishable trailing, leading, or parallel stratiform

regions with respect to the propagating convection line (e.g., Houze et al. 1990; Parker and Johnson 2000) with distinctive kinematic, thermodynamic, and microphysical properties (e.g., Zipser 1977; Houze 1977, 1989), making them particularly amenable to observational and modeling studies. Squall lines commonly rely on convective downdrafts driven by evaporating condensate to transport horizontal momentum (e.g., LeMone 1983; LeMone et al. 1984) and relatively drier, lower equivalent potential temperature midlevel air to lower levels, forming a cold pool that often propagates outward (e.g., Braham 1952; Fujita 1959; Zipser 1969; Moncrieff and Miller 1976). The edge of the cold pool often acts as a focus for subsequent deep convection should environmental stability, humidity, and vertical wind shear allow it (e.g., Newton 1950;

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Houze and Betts 1981; Thorpe et al. 1982; Weisman et al. 1988; Skamarock et al. 1994). A maturing squall line often forms organized mesoscale circulations in response to sustained deep convection that further impact the system evolution (e.g., Houze 1977; Zipser 1977; Houze and Betts 1981; Gamache and Houze 1982; Smull and Houze 1985, 1987).

Cloud system resolving models with horizontal grid spacing of approximately 0.5 to 5 km have been used for decades to help researchers understand the life cycle of different types of squall lines and their sensitivity to environmental conditions. Many studies have shown that models readily produce squall lines given favorable environmental conditions and are capable of reproducing characteristic tropical and midlatitude squall-line features including the following: rear inflow and front-to-rear (storm relative) flow with distinctive mesoscale updrafts and downdrafts (e.g., Ogura and Liou 1980; Rutledge and Houze 1987; Lafore and Moncrieff 1989; Weisman 1992); bow echoes (e.g., Weisman 1993; Przybylinski 1995); line-end vortices (e.g., Weisman and Davis 1998; Weisman and Trapp 2003; Trapp and Weisman 2003); and sensitivity of system longevity and propagation to ambient vertical wind shear (e.g., Fovell and Ogura 1988; Rotunno et al. 1988; Weisman and Rotunno 2004; Coniglio et al. 2006).

Despite all of the successes of past observational and modeling research that have led to the current advanced understanding of squall lines, a remaining question is how well cloud system resolving models reproduce real world sensitivities of squall-line features to environmental thermodynamic and kinematic conditions. Are these models prone to over or undersimulating squall-line modes in certain environments? Are they biased in representing squall-line precipitation evolution or propagation? These questions have become increasingly important because large-domain cloud system resolving models are now being used in short-range weather forecasting (e.g., Lean et al. 2008; Weisman et al. 2008; Clark et al. 2011; Benjamin et al. 2016; Powers et al. 2017). While these models are certainly an improvement over their coarser-resolution predecessors, some model biases still exist at these scales that impact weather forecasts. They are also important from a climate prediction standpoint since regional climate models are now being used to anticipate how “extreme” weather such as severe deep convective events will change with the climate (e.g., Prein et al. 2017a,b; Rasmussen et al. 2017).

Numerous cloud system resolving intercomparison studies have highlighted significant sensitivity of simulated mesoscale convective system (MCS) cloud and precipitation evolution to the choice of physics

parameterization, the microphysics scheme in particular (e.g., Redelsperger et al. 2000; Xu et al. 2002; Varble et al. 2011, 2014a,b; Fridlind et al. 2012; Zhu et al. 2012; Morrison et al. 2015a,b; Fan et al. 2017; Xue et al. 2017; Han et al. 2019). Despite significant sensitivities of simulations to microphysics parameterizations, several model biases have been shown to be persistent across a range of model setups at cloud system resolving scales. Cloud system resolving models tend to overpredict Rayleigh radar reflectivity because of excessive production of large ice, particularly in the form of graupel or hail (e.g., Blossey et al. 2007; Matsui et al. 2009; Lang et al. 2011; Varble et al. 2011, 2014a; Hagos et al. 2014; Fan et al. 2017; Stanford et al. 2017), and recent studies have shown that they often produce convective updrafts that are too strong (Varble et al. 2014a; Fan et al. 2017). These deep convective biases can be partially alleviated through decreasing model grid spacing, which results in smaller convective drafts that promote more efficient precipitation of rimed ice out of updrafts (Varble et al. 2014a).

Model microphysical biases are not confined to deep convective regions of MCSs. For example, recent studies show that cloud-system resolving simulations commonly produce insufficient stratiform precipitation with respect to observations (e.g., Li et al. 2009; Morrison et al. 2009; Luo et al. 2010; Varble et al. 2011, 2014b; Hagos et al. 2014; Tao et al. 2016), and several studies have shown that simulations often struggle to produce robust transition regions (e.g., Biggerstaff and Houze 1991, 1993; Braun and Houze 1994) between convective and stratiform regions in which a localized radar reflectivity and rain rate minimum is observed (Morrison et al. 2015a; Fan et al. 2017; Jensen et al. 2018; Tridon et al. 2019). Han et al. (2019) showed that spread in simulated stratiform rainfall was correlated with differences in simulated deep convective detrainment of ice. Therefore, it is reasonable to suspect that biased deep convective kinematic and/or microphysical properties may bias overall system evolution. Several recent studies show differences between observed and simulated MCS life cycles and geographical distribution over the U.S. Great Plains for seasonal and multiyear simulations using horizontal grid spacing of 3–4 km (Prein et al. 2017c; Feng et al. 2018; Haberlie and Ashley 2019; Scaff et al. 2019). However, the ways in which underresolved deep convective drafts within MCSs impact mesoscale kinematic and microphysical evolution remain unclear. Weisman et al. (1997) show that MCSs evolved more slowly and became larger in 4-km runs than 1-km runs because of changes to vertical mass transport and cold pool formation, while Schwartz et al. (2009) show little difference between

2- and 4-km convective forecasts. Other studies have shown that simulations with 1–3 km grid spacing have larger cold pool property biases than higher-resolution simulations that may impact MCS evolution (e.g., Dawson et al. 2010).

There is considerable evidence that individual deep convective drafts are underresolved in models with horizontal grid spacing larger than a few hundred meters. Several studies have shown that simulated and theoretical convective drafts often transition from plume-like to thermal-like structures moving from 1000 to 100 m horizontal grid spacing with significant impacts on convective draft and system properties (e.g., Petch et al. 2002; Bryan et al. 2003; Craig and Dörnbrack 2008; Varble et al. 2014a; Bryan and Morrison 2012). Lebo and Morrison (2015) showed that the peak of the vertical velocity kinetic energy spectra associated with the mean convective updraft size in a squall-line simulation became resolved as the horizontal grid spacing was reduced to 0.25 km and below. This also corresponded to a transition in the flow from a laminar to turbulent regime, consistent with Bryan et al. (2003). A significant reduction in convective mass flux also occurred moving from 0.5 to 0.25 km grid spacing, with limited sensitivity for coarser grid spacings. The convective core number increased and mean area decreased as grid spacing reduced from 2 to 0.25 km but remained fairly steady as grid spacing was reduced further below 0.25 km. Jeevanjee (2017) presented further theoretical support for 250 m being the coarsest grid spacing at which convection begins to be resolved. However, these results are dependent on the model setup and meteorological conditions as evidenced by a lack of convergence in some convective core statistics in some other studies (e.g., Bryan et al. 2003; Morrison et al. 2015b). Details aside, a typical convective updraft that is ~2 km wide can only be resolved with a grid spacing of less than 250–400 m in most atmospheric models that typically have effective resolutions that are 5–7 times the grid spacing (e.g., Skamarock 2004).

Although it is well established that horizontal grid spacing of a few hundred meters or less is needed to resolve deep convective drafts, the impacts of under-resolving drafts on MCS structure and behavior in kilometer-scale simulations are unclear. We evaluate the representation of precipitation structure in cloud system resolving simulations of the 20 May 2011 squall-line event during the Mid-Latitude Continental Convective Clouds Experiment (MC3E; Jensen et al. 2016) in Oklahoma using radar observations. This case has been previously simulated in several previous studies (Tao et al. 2013; Fan et al. 2015; Marinescu et al. 2016; Saleeby et al. 2016; Tao et al. 2016; Fan et al. 2017;

Fridlind et al. 2017; Xue et al. 2017; Cheng and Zhang 2019; Han et al. 2019; Stanford et al. 2019). Biases in simulated precipitation structure are then connected to differences in observed and simulated mesoscale circulations. Sensitivities of simulated squall-line circulation and precipitation feature biases to convective updraft and downdraft properties early in the squall-line life cycle are then analyzed by comparing simulations with horizontal grid spacings of 750 and 250 m. Section 2 highlights datasets and methods used, section 3 shows comparisons between simulations and observations, section 4 investigates simulated squall-line thermodynamic and kinematic evolution, section 5 links simulated squall-line evolution to convective draft properties, and section 6 summarizes primary conclusions.

2. Data and methods

a. Squall-line event

Early on 20 May 2011 during the MC3E field campaign, a cold front associated with a weak low pressure center in western Kansas (Fig. 1b) moved through the panhandles of Texas and Oklahoma. The developing frontal cyclone was associated with an upper-level trough centered over Wyoming (Fig. 1a) moving northeastward in time. When the cold front passed a dryline near the border of Oklahoma and the Texas panhandle, deep convection began initiating along it by 0500 UTC. This deep convection continued strengthening and moved eastward as it formed into a mature squall line with a trailing stratiform region extending from northern Texas to the Kansas–Oklahoma border by 1000 UTC (Fig. 1c). After 1000 UTC, the deep convection weakened considerably and became more isolated as the convective region became detached from the stratiform region at low levels. The focus of this study is on the early to mature life cycle stages of the squall line between 0430 and 1000 UTC.

b. Model setup

Simulations are performed using the Weather Research and Forecasting (WRF) Model V3.8.1 (Skamarock and Klemp 2008; Skamarock et al. 2008) with 6-hourly boundary conditions provided by the National Centers for Environmental Prediction Final (FNL) global operational analyses (NOAA/National Centers for Environmental Prediction 2000). The simulations are initialized at 0000 UTC 20 May 2011. Two-way nesting is used on 4 domains with horizontal grid spacing of 20.25, 6.75, 2.25, and 0.75 km for the coarse-resolution run centered on southwestern Oklahoma (Fig. 2). An additional 0.25 km horizontal grid spacing domain is

FIG. 1. The large-scale meteorological setup just after initial deep convective initiation along the front and dryline boundary near the southwestern corner of Oklahoma showing (a) 500-mb wind speed (filled), wind vectors, and geopotential height (contoured every 30 m), and (b) 2-m moist static energy (filled), wind

vectors, and surface pressure (contoured every 2 mb). Analyses are derived from NCEP FNL at 0600 UTC 20 May 2011, and (c) 2.5-km altitude Rayleigh reflectivity observed by NEXRAD at 1000 UTC.

added for the high-resolution simulation with the expectation that typical convective updrafts will be resolved at this grid spacing and not resolved using 750-m grid spacing following the results of previous literature highlighted in the introduction. The domain with 750-m grid spacing is 594 km $\times$ 594 km while the 250-m grid spacing domain is 315 km $\times$ 315 km. 100 vertical half (mass) levels are used with a model top near 20.5-km altitude, varying the eta value 0.01 between each level. This yields grid spacing that increases from $\sim$ 85 m near the surface to $\sim$ 100 m at 2.5-km altitude, 200 m at 9.5-km altitude, and 400 m at 15-km altitude. Sensitivity to vertical resolution was not tested. Time steps of 2 and 2/3 s are used in 750- and 250-m domains, respectively.

The Kain–Fritsch cumulus parameterization (Kain 2004) is used in the 2 outermost domains. The Eta similarity surface layer scheme (Monin and Obukhov 1954; Janjić 1994, 1996, 2002) is used with the Noah land surface model (Tewari et al. 2004). The Rapid Radiative Transfer Model longwave (Mlawer et al. 1997) and Dudhia shortwave (Dudhia 1989) radiation parameterization are employed. The Morrison microphysics scheme (Morrison et al. 2009) is used with hail-like rather than graupel-like properties for the dense precipitating ice category. The Mellor–Yamada–Janjić planetary boundary layer scheme (Janjić 1994) is used in the 4 outermost domains including the 750-m grid spacing domain, and horizontal diffusion is diagnosed from horizontal deformation following Smagorinsky (1963). No boundary layer scheme is used in the 250-m grid spacing domain. Instead, diffusion is handled using a prognostic equation for three-dimensional turbulent kinetic energy. Differing mixing parameterizations and time steps may contribute to differences in the atmospheric evolution, notably near the surface, although impacts on the early life cycle of the squall line may be mitigated by the nocturnal initiation along the cold front. As the squall line grows, it also influences the environmental conditions around it, and this could also contribute to some differences in the evolution of the squall line in each simulation.

Simulated Rayleigh radar reflectivity (Z_e) that is used for model–observation comparisons is computed by summing reflectivity contributions from 3 hydrometeor species: rain, snow, and hail [Eq. (1)]. Because precipitation echoes are the focus in this study, reflectivity components from cloud ice and water that

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FIG. 2. (a) Domains used for nested WRF simulations in orange with NEXRAD locations and range rings in black, NOAA radiosonde locations in light blue, and ARM observing sites in dark blue. The red box shows the 250-m grid spacing domain. (b) As in (a), but zoomed in on the 750-m domain.

are quite small are neglected. For each species i , N_i is the particle number concentration, λ_i is the slope of the particle gamma size distribution and a function of predicted number and mass mixing ratios, α_i and β_i are coefficients of the particle mass–size relationship ($m = \alpha D^\beta$), ρ_w is the density of liquid water, and μ_i is the shape parameter of the particle gamma size distribution. In the Morrison microphysics scheme, $\mu = 0$ for each precipitation species: rain, snow, and hail. For ice species, Eq. (1) is additionally multiplied by a factor of 0.224, which accounts for the different dielectric factors for liquid and ice, following Smith (1984):

$$Z_e = 1 \times 10^{18} \sum_{i=1}^3 N_i \lambda_i^{-2\beta} \left[\frac{6\alpha_i}{\pi\rho_w} \right]^2 \frac{\Gamma(2\beta_i + \mu_i + 1)}{\Gamma(\mu_i + 1)}. \quad (1)$$

Simulated Rayleigh radial velocity is also calculated for the KFDR, KTLX, and KVN NOAA NEXRAD radar locations by computing the dot product between the Rayleigh-weighted hydrometeor movement vector and the unit vector describing the direction from the radar location to the model grid point of interest. The locations of these radars are shown in Fig. 2. The Rayleigh-weighted hydrometeor movement vector has zonal wind, meridional wind, and vertical Doppler velocity components. The vertical Doppler velocity is a combination of the vertical wind speed (w) and the Rayleigh-weighted hydrometeor fall speed [Eq. (2)]. Cloud ice and water fall speeds are assumed to be negligible relative to precipitation fall speeds and wind speeds. Therefore, the Rayleigh-weighted hydrometeor fall speed has contributions from three hydrometeor species (rain, snow, and hail) where $v_i(D)$ is the particle terminal fall speed relationship and $N_i(D)$ is the particle size distribution for species i :

$$V_{\text{Dopp}} = w - \frac{\sum_{i=1}^3 \int_0^\infty v_i(D) D^6 N_i(D) dD}{\sum_{i=1}^3 \int_0^\infty D^6 N_i(D) dD}. \quad (2)$$

All comparisons between the two simulations in this study are performed on a common 750-m grid. For the 250-m run, this was accomplished by using output from the 750-m grid spacing parent domain that fit within the 250-m domain lateral boundaries. WRF linearly averages finer-resolution nests to the coarser parent grid such that the 250-m run output used for all comparisons is effectively averaged to the 750-m grid spacing.

c. Observations

NOAA NEXRAD Level II data (NOAA/NWS Radar Operations Center 1991) from the KFDR, KTLX, and KVN radars in Oklahoma are used for performing comparisons of observed and simulated radar reflectivity and radial velocity. Radial velocities are dealiased using the region-based dealiasing algorithm within the Python Atmospheric Radiation Measurement (ARM) Radar Toolkit (Py-ART, Helmus and Collis 2016), and radar data are converted from polar to Cartesian grids using NCAR RadX software (<https://www.eol.ucar.edu/content/lidar-radar-open-software-environment>). For domain-wide radar reflectivity fields, radar data are merged by applying measurements from the nearest radar to each grid point.

NOAA radiosonde observations from the Integrated Global Radiosonde Archive (IGRA) Version 2 (Durre et al. 2016) and U.S. Department of Energy ARM radiosonde observations (Coulter et al. 1994) surrounding

FIG. 3. Soundings on skew T -log p diagrams from (a) Amarillo, TX, at 0000 UTC, (b) Fort Worth, TX, at 0000 UTC, (c) Vici, OK, at 0300 UTC, and (d) Purcell, OK, at 0600 UTC. Observations are shown in thin black and the model output is shown in thick black (smoother profiles and rightmost wind barbs).

the 250-m domain are used to evaluate the model initialization after interpolation to model levels.

3. Comparison of simulations with observations

If a simulation is to properly reproduce the observed mesoscale kinematic and microphysical evolution of a squall line, the large-scale kinematic and thermodynamic conditions need to be properly initialized. NOAA

radiosonde measurements from the U.S. National Weather Service (light blue diamonds in Fig. 2) and ARM radiosondes (blue diamonds in Fig. 2) are used to evaluate the accuracy of the model initialization at 0000 UTC 20 May 2011 in Fig. 3. The Amarillo, Texas, sounding shows that the thermodynamic and kinematic conditions in the western part of the domain are fairly well represented with temperature, humidity, wind speed, and wind direction profiles that only slightly

differ and boundary layer depths that are the same. The Fort Worth, TX soundings shows that temperature, humidity, wind speed, and wind direction profiles are also fairly comparable in observations and simulations on the eastern edge of the 250-m domain. This good agreement is likely partly due to assimilation of NOAA radiosonde observations into FNL analyses used to initialize WRF. However, the unassimilated ARM soundings (0300 UTC at Vici, Oklahoma, and 0600 UTC at Purcell, Oklahoma) also generally agree with WRF profiles in the free troposphere. The WRF initialization appears to have a lower boundary layer moist bias ahead of the dryline and cold front. This causes higher than observed surface-based convective available potential energy (CAPE), although mixed layer CAPE is not significantly higher in the model than observations. Water vapor and wind magnitudes in the free troposphere also differ slightly, more for the ARM soundings than the NOAA soundings. In addition, the model representation of the dryline and cold front structures is unknown, although the initiation of deep convection in space and time is similar to that observed between 34° and 35° latitude where the squall line early life cycle is analyzed.

After the squall line develops, it moves eastward in observations and simulations. Using the 2.5-km altitude 40 dBZ echo between 34° and 36°N latitude as a proxy for the leading edge of the squall line shows that the simulated squall lines move faster than observed between 0600 and 0800 UTC, but between 0800 and 1000 UTC the observed squall line partially catches up to the simulated lines (Fig. 4). The simulated squall movement is faster in the 750-m run than the 250-m run. Additionally, the simulated squall line in the 750-m run develops a notable bowing segment that is not present in observations or the 250-m run to the same degree (e.g., 0800 UTC in Fig. 3).

Although the simulations reproduce observed large-scale thermodynamic and kinematic conditions with an eastward moving squall line of similar extent, comparisons of radar Rayleigh reflectivity structure in Fig. 5 show considerable differences. Both simulations have wider regions of 2.5-km altitude reflectivity greater than 40 dBZ than observed between 0700 and 0900 UTC. These high reflectivity regions are associated with the leading deep convective line. The 750-m simulated high reflectivity convective region is wider than in the 250-m simulation (Fig. 5). Both simulations form a low reflectivity “transition” region between the convective and stratiform regions like observed for this and many other cases (e.g., Ligda 1956; Houze 1977; Smull and Houze 1985; Srivastava et al. 1986; Roux 1988; Rutledge et al. 1988; Biggerstaff and Houze 1993; Braun and Houze 1994). The

FIG. 4. Locations of the leading edge of the squall line in observations (grayish colors), the 750-m grid spacing simulation (reddish colors), and the 250-m grid spacing simulation (bluish colors) at 0600 (lightest colors), 0800, and 1000 (darkest colors) UTC.

stratiform low level reflectivity is greater in simulations than observed, most notably in the 750-m simulation, which has a south-southwest (SSW)–north-northeast (NNE)-oriented strip of enhanced reflectivity reaching 45 dBZ at 0900 UTC that is more muted in the 250-m simulation and not present in observations.

There are also differences in the vertical structure of radar reflectivity among the simulations and observations (Fig. 6). The leading edge of the observed high reflectivity convective region is nearly vertical at 0800 UTC. This suggests that the observed convective updrafts are not significantly sheared in the zonal direction. The simulations do not reproduce this feature, instead exhibiting a convective region reflectivity structure that tilts upshear from the leading squall-line edge with peak reflectivities at all altitudes occurring about 15 km behind the leading squall-line edge (defining the “edge” by location of the 40 dBZ echo at 2.5-km altitude at each latitude and averaging each latitude based on distance from the edge rather than longitude). Simulations also produce noticeably wider regions of reflectivity >40 dBZ than observed. For example, at 0800 UTC the observed 40 dBZ region at 2.5-km altitude is 25 km wide, whereas it is 45 km wide in the 250-m run and 50 km wide in the 750-m run. Although the magnitude of this difference varies in time, it is always present between 0700 and 1000 UTC with the observed 40+ dBZ region never exceeding 30 km in width and simulated 40+ dBZ regions almost always exceeding 40 km in width (not shown).

Radial velocity observations from the KTLX radar show that model-observation differences are not limited

FIG. 5. 2.5-km altitude Rayleigh radar reflectivity (a)–(c) observed by NEXRAD, (d)–(f) simulated in the 750-m run, and (g)–(i) simulated in the 250-m run at (a),(d),(g) 0700, (b),(e),(h) 0800, and (c),(f),(i) 0900 UTC. Black contours signify the leading edge of the squall line. The regions to the west of these edges bounded by the constant latitude black lines are the regions used for computing average vertical cross sections shown in Fig. 6.

to precipitation structure. Figure 7 shows westward viewing vertical cross sections of KTLX-observed radial velocity at 1000 UTC and simulated radial velocity at 0930 UTC when the simulated squall line is at a similar location as observed to the west of KTLX. The rear inflow jet and front-to-rear (FTR) flow are clearly visible in all cross sections; however, the location and strengths

of these circulations varies between observations and simulations. The observed rear inflow jet in the zonal direction consistently exceeds 25 m s^{-1} in strength with maximum strength generally located between 2 and 3 km altitude. The 250-m run produces a jet of similar altitude that is perhaps slightly weaker while the 750-m run produces a noticeably weaker jet peaking in strength

FIG. 6. Squall-line Rayleigh reflectivity vertical cross sections at 0800 UTC averaged over the latitudes shown in Fig. 4 with respect to the leading edge of the convective region defined using the farthest eastward position of the 40 dBZ echo at a given latitude. (a) NEXRAD observations, (b) the 750-m simulation, and (c) the 250-m simulation are shown.

between 1.5 and 3 km altitude. This difference in rear inflow altitude between the simulations exists for most locations and times (not shown). The weaker jet in the 750-m run is associated with a weaker, less concentrated negative pressure perturbation beneath the tilted convective updrafts than in the 250-m run (not shown), likely related to the more tilted, weaker updrafts at this time in the 750-m run. The farther descent of the rear inflow in the 750-m may cause additional weakening of the wind speed relative to the 250-m run.

Rear inflow differences are also correlated with FTR flow strength and altitude differences. Observed FTR flow exceeds 15 m s^{-1} at times (as in Fig. 7) and increases in altitude moving back from the edge of the squall line. Observed flow is generally westward above 5 km altitude and eastward below this level. In the 750-m run, this delimiter is 3-km altitude and in the 250-m run, it is 4-km altitude. 250-m FTR strength also exceeds that in the

750-m run in better agreement with observations. Although Fig. 7 shows a single radar (KTLX) at a single time during the mature stage of the squall line, it is representative of differences throughout the squall line (not shown). Therefore, neither simulation reproduces observed mesoscale zonal circulations in the squall line, but the 250-m run produces circulations that are closer to those observed.

4. Contrasting thermodynamic and kinematic evolution of the simulated squall lines

Both simulations fail to reproduce aspects of the observed squall-line horizontal reflectivity structure. This may partly be a result of differences in cold front and dryline structure in reality compared to the analyses used to force the simulations. However, the 250-m run better reproduces the observed moving speed.

FIG. 7. NOAA KTLX radar west-east vertical cross sections showing radial velocity (filled; black contours at -15 , -5 , 5 , 15 , and 25 m s^{-1}) and reflectivity contoured at 5 (green) and 45 (orange) dBZ to highlight convective and stratiform precipitation regions. (a) Observations at 1000 UTC, (b) the 750-m simulation at 0930 UTC, and (c) the 250-m simulation at 0930 UTC are shown.

Moreover, the 250-m run better reproduces observed radar reflectivity and Doppler velocity fields associated with convective precipitation, the rear inflow jet, and the front-to-rear flow. The possible causes of these differences are investigated further in this section.

The rear inflow and FTR circulation differences between the 750- and 250-m simulations are associated with buoyancy structure differences that trace back to the early stages of the squall-line development. The composite squall-line negative buoyancy perturbation associated with the cold pool is greater at 0700 UTC in the 250-m run than the 750-m run (Figs. 8e,f) where the buoyancy includes temperature, moisture, and condensate contributions (pressure contribution is neglected) and is computed as a perturbation with respect to the domain horizontal mean at each constant altitude. In addition, the leading edge of the propagating cold pool

in the 250-m run at 0700 UTC is far more upright than the significantly upshear-tilted cold pool within the 750-m run. The tilt of the cold pool leading edge in the 750-m run is associated with the same structure in the zonal wind field that highlights the descent of the rear inflow to near the surface. In contrast, the rear inflow remains more elevated in the 250-m run, providing a positive vorticity circulation at low levels that helps to counteract the negative vorticity circulation of the cold pool (Fig. 9). Both simulations exhibit significant upshear tilting of updrafts associated with positive ambient vorticity that is weaker in magnitude than the negative vorticity associated with the cold pool and rear inflow circulations (Fig. 9). However, the strong negative vorticity region along the upper head of the cold pool and rear inflow jet is lower in altitude and far more upshear tilted in the 750-m run than the 250-m run,

FIG. 8. Squall-line zonal wind (color fill) and buoyancy (contoured every 0.04 m s^{-2} with negative values in blues and positive values in oranges) vertical cross sections averaged between 34° and 35°N latitude at (a),(b) 0500, (c),(d) 0600, and (e),(f) 0700 UTC for the (a),(c),(e) 750-m and (b),(d),(f) 250-m simulations.

which correlates with significantly greater upshear tilting of updrafts (Fig. 9). This is consistent with Rotunno–Klemp–Weisman (RKW) theory (Rotunno et al. 1988; Weisman and Rotunno 2004) in which the more upright

nature of the cold pool leading edge in the 250-m run promotes a more upright positive buoyancy perturbation associated with the squall-line deep convection (Fig. 8).

FIG. 9. Squall-line meridional vorticity (color fill), vertical motion [contours at -0.5 m s^{-1} (cyan) and 1, 2, and 4 m s^{-1} (oranges)], and zonal wind (dashed gray contours at -6 and -2 m s^{-1} and solid gray contours at 2, 6, 10, 14, 18, and 22 m s^{-1}) vertical cross sections averaged between 34° and 35°N latitude at 0700 UTC for the (a) 750-m and (b) 250-m simulations.

FIG. 10. Squall-line zonal wind tendency due to vertical advection (color fill), vertical motion [contours at -0.5 , -1 , and -2 m s^{-1} (cyan) and 1 , 2 , and 4 m s^{-1} (oranges)], and zonal wind (dashed gray contours at -6 and -2 m s^{-1} and solid gray contours at 2 , 6 , 10 , 14 , 18 , and 22 m s^{-1}) vertical cross sections averaged between 34° and 35°N latitude at 0520 UTC for the (a) 750-m and (b) 250-m simulations.

Tracing the evolution of the buoyancy perturbation and zonal wind fields back in time to 0500 and 0600 UTC indicates westerly zonal momentum being more effectively transported downward behind the leading squall-line edge in the 750-m run than the 250-m run (Figs. 8a–d). This is more supported in Fig. 10 by noticeably greater downward advection of zonal wind by convective downdrafts in the 750-m run in the region where relatively greater zonal winds extend down to the surface (Fig. 10). This is associated with a cold pool that quickly becomes upshear tilted, promoting more upshear-tilted deep convection in the 750-m run than the 250-m run. As early as 0500 UTC before the squall line exists, when initial deep convection is forming, the cold pool and downward transport of zonal momentum is already greater in the 750-m than 250-m run. This difference is associated with differing kinematic, thermodynamic, and microphysical structures in the initial deep convection that forms in the two simulations, which is discussed next.

At 0500 UTC during the early initiating stages of deep convection the 750-m run already has more downward motion and associated latent cooling behind the leading edge of the squall line than the 250-m run (Figs. 11a,b). By 0600 UTC, the upward motion and latent heating are noticeably more upshear tilted in the 750-m run with peak latent cooling 10–50 km rearward of the squall-line leading edge and peak downward motion nearly 30 km rearward. In contrast, peak latent cooling in the 250-m run is 5–25 km rearward of the leading squall-line edge with peak downward motion 10 km rearward. These differences are correlated with condensate contents exceeding 1.6 g m^{-3} in the ice region that extend 35 km to rearward of the squall-line edge in the 750-m run but only 25 km rearward in the 250-m run. This highlights how the more upshear-tilted convective updrafts in the 750-m run transport more condensate rearward that falls

into drier air and causes greater latent cooling and downward motion. The greater latent cooling and downward motion can then create a situation in which the rear inflow descends farther toward the surface and a surging cold pool can be created that further reinforces the upshear tilt of the deep convection in a positive feedback as long as the leading line convection remains strong enough to detrain sufficient condensate mass. Detrainment of condensate from convective updrafts is a critical component of this cycle because its evaporation in subsaturated air creates cooling that is a primary driver of downdrafts that transport momentum and colder, drier air downward where they impact the cold pool and rear inflow evolution (see references in section 1). For this case, the convection weakens in the 750-m run compared to the 250-m run, which is apparent by 0700 UTC in Fig. 11, likely caused by weaker lift associated with the shallower, more upshear-tilted cold pool relative to the 250-m run. The 250-m run and observations also eventually evolve to this state of the cold pool circulation becoming stronger than the presquall vertical wind shear circulation such that lifting along the gust front weakens and becomes too shallow to sustain the squall line, but they do so later than occurs in the 750-m run.

5. Comparison of simulated convective updraft and downdraft properties

Differences in the kinematic and thermodynamic characteristics of the initial simulated deep convection may be caused by differences in the initial deep convective updraft and downdraft properties. To analyze simulated updraft and downdraft properties throughout this section, updrafts and downdrafts are defined separately on constant altitude levels as contiguous grid points with condensate mass greater than 0.1 g m^{-3} and

FIG. 11. Squall-line latent heating (color fill), vertical motion [contoured at -0.5 , -1 , and -2 m s^{-1} (blues) and 1 , 2 , 4 , and 8 m s^{-1} (green to orange)], and total condensed water content (contoured in gray at 0.1 , 0.2 , 0.4 , 0.8 , 1.6 , and 3.2 g m^{-3}) vertical cross sections averaged between 34° and 35°N latitude at (a),(b) 0500, (c),(d) 0600, and (e),(f) 0700 UTC for the (a),(c),(e) 750-m and (b),(d),(f) 250-m simulations.

vertical velocity above 2 m s^{-1} and below 2 m s^{-1} , respectively. Total downdraft mass flux averaged over the first two hours of deep convective activity is 50% greater in the 750-m run than the 250-m run at 2-km altitude, and significant differences extend throughout the troposphere (Fig. 12a). This is associated with greater downdraft latent cooling throughout the troposphere in the 750-m run as well (Fig. 12b). The updraft mass flux and latent heating is also significantly greater in the 750-m run than the 250-m run above 1.5-km altitude. Below 1.5-km altitude, the greater downdraft cooling and mass flux in the 750-m run exceeds updraft heating and mass flux differences. Hence, the 750-m run has greater net convective cooling and downward motion at low levels than the 250-m run. Above 1.5-km altitude, heating and upward mass flux differences between the two simulations exceed cooling and downward mass flux differences, respectively. Therefore, net convective heating and upward mass flux are greater in the 750-m run than the 250-m run at these levels.

Differences in convective draft kinematic and thermodynamic properties are correlated with differences in

microphysical properties. The 750-m simulation has more total condensate in both convective updrafts and downdrafts (Fig. 13), with nearly double the updraft condensate and 40% more downdraft condensate at 2-km altitude. Most of these differences in condensate mass are from differences in graupel between 4- and 10-km altitudes. Cloud water and rain both contribute strongly to simulated updraft differences below 4-km altitude, while rain differences dominate in downdrafts. Although not shown, downdraft relative humidity differences are not significantly different between the two simulations, indicating that a primary driver of the greater overall downdraft cooling and mass flux is greater downdraft condensate that may originate from detrainment of greater updraft condensate mass.

Greater condensate mass is partitioned among fewer updrafts and downdrafts in the 750-m run than the 250-m run (Fig. 14a; recall that 250-m run drafts are averaged to 750-m grid spacing to match the 750-m run). Above 3-km altitude, the 250-m run has approximately 25%–100% greater convective draft numbers, depending on altitude. This difference is more than offset by

FIG. 12. Total updraft and downdraft (a) mass flux and (b) latent heating/cooling profiles averaged from 0400 to 0600 UTC between 34° and 35°N latitude in the 750-m (red) and 250-m (blue) simulations. Simulation differences are consistent with differences over the entire domain. Updraft and downdraft objects in this figure and figures to follow are identified as contiguous grid points at each constant altitude level with condensate mass above 0.1 g m^{-3} and vertical velocity above 2 m s^{-1} and below 2 m s^{-1} , respectively.

750-m run updrafts and downdrafts that are significantly larger than 250-m drafts at all altitudes (Fig. 14b). Average 750-m run downdraft areas are roughly 1.5–2 times larger than average 250-m run downdrafts while average updrafts are roughly 2–2.5 larger depending on altitude. These significant changes in updraft size between the 750- and 250-m grid spacings are consistent with Lebo and Morrison (2015). The 750-m run has

greater updraft and downdraft total areal coverages than the 250-m run, consistent with greater differences in average draft area than number of drafts between the runs.

Downdraft condensate mass and latent cooling increase as downdraft area increases while average and peak vertical velocity decrease with increasing area. The sensitivities of these thermodynamic, kinematic, and microphysical variables to downdraft area are very

FIG. 13. (a) Updraft and (b) downdraft total condensate averaged from 0400 to 0600 UTC between 34° and 35°N latitude for the 750-m (solid) and 250-m (dashed) simulations. Total condensate is shown in black, rain in blue, cloud water in cyan, cloud ice/snow in orange, and graupel in red. Simulation differences are consistent with differences over the entire domain.

FIG. 14. Profiles of the total (a) number of updrafts (solid) and downdrafts (dashed) and (b) their mean area averaged from 0400 to 0600 UTC between 34° and 35°N latitude in the 750-m (red) and 250-m (blue) simulations.

similar in both simulations (Fig. 15). Therefore, greater downdraft condensate mass and latent cooling in the 750-m run than the 250-m run highlighted previously and by the symbols in Fig. 15 do not result from differences in thermodynamic, microphysical, or kinematic properties of a given downdraft size. They instead result from a difference in the downdraft size distribution with the 750-m run having less relatively small downdrafts and more numerous relatively large downdrafts (greater than $\sim 1 \text{ km}^2$ in area). This difference in the downdraft size distribution also causes 750-m run downdrafts to exhibit stronger mean and peak downward motion (Figs. 15a,b), which can be expected to result in greater downward penetration of the downdrafts.

The same result applies to convective updrafts in which updraft thermodynamic, microphysical, and kinematic properties as a function of updraft area are similar in both simulations (Fig. 16). As updraft area increases, mean and maximum vertical velocity, condensed water content, and latent heating all increase consistent with previous studies (e.g., Khairoutdinov and Randall 2006; Kirshbaum and Grant 2012; Morrison 2016a,b, 2017; Rousseau-Rizzi et al. 2017), and the 750-m run has a greater number of relatively large updrafts than the 250-m run. Mean and maximum vertical wind speed are actually greater for a given updraft area in the 250-m run, but the difference in the simulated updraft size distributions outweighs this effect to produce median and mean updraft vertical winds that are greater in the 750-m run (Figs. 16a,b).

It is likely that relatively larger downdrafts in the 750-m run would result even without the larger updrafts.

However, Lucas et al. (1994) show using aircraft measurements that updraft and downdraft widths vary similarly as environmental conditions change. Additionally, downdrafts and the evaporating condensate that fuel them result from condensate that is detrained from the updrafts that produce the condensate. Therefore, it is reasonable to suspect that relatively larger updrafts with greater condensed water amounts aid in the production of larger regions of detrained condensate that help to fuel larger downdrafts which can penetrate farther downward because of greater capacity for latent cooling. Marion and Trapp (2019) present evidence for this explanation based on supercell simulations. There are also possible dynamical reasons for downdraft size being linked to updraft size. Morrison (2016b) shows that wider updrafts cause wider regions of downward directed buoyant pressure forces that could force wider downdrafts. Additionally, wider updrafts may be associated with wider regions of downward dynamic pressure forcing and lower-frequency gravity waves that could cause wider downdrafts, although analyzing these connections is beyond the scope of this study. Last, the cores of wider convective drafts are likely to be more slowly diluted from turbulent entrainment which retains greater magnitudes of buoyancy and vertical motion (e.g., Morrison 2017), enhancing the likelihood of greater vertical penetration by drafts. However, Lebo (2018) also shows that updraft velocity changes slowly as updraft width increases if updrafts are highly sloped, which may limit the effects of entrainment in this case.

The connection between updraft and downdraft properties can be seen in their changes as a function of time (Fig. 17). Updraft and downdraft number are

FIG. 15. 2.5-km altitude mean downdraft (a) mean vertical wind speed, (b) minimum vertical wind speed, (c) condensed water content, and (d) latent cooling as a function of downdraft horizontal area between 0400 and 0600 UTC from 34° to 35°N latitude (solid lines). The downdraft area distributions for each simulation are dashed. Distributions are computed by averaging each downdraft variable (mean vertical wind speed, minimum vertical wind speed, condensed water content, and latent cooling) binned by downdraft area with bin sizes equaling the size of a single grid box (0.5626 km^2). Squares represent mean values and diamonds represent median values of all downdraft samples. The 750-m run is shown in red and the 250-m run in blue.

correlated without any time lag (Fig. 17b) as are updraft and downdraft mean areas (Fig. 17c) highlighting the linkage between the two. Interestingly, the number of drafts remains relatively constant between 0400 and 0530 UTC with similar numbers of updrafts and downdrafts within each simulation. During this time period, the mean draft areas increases substantially. After 0530 UTC, the mean individual draft areas level out and eventually begin to decrease but the number of drafts rises dramatically during this period. Total updraft and downdraft mass fluxes steadily increase from 0430 UTC onward indicating the increases in mass fluxes are initially driven by increasing draft sizes and thereafter driven by increasing draft numbers. This could potentially be related to the cold pool, which is well defined at 0600 UTC but not at 0500 UTC (Fig. 8), becoming the primary feature initiating and impacting the shearing of new updrafts.

Updraft and downdraft total mass fluxes are also correlated in time (Fig. 17a) with differences between the two simulations driven more by differences in total updraft and downdraft areal coverage than differences in draft vertical wind speed (not shown). These results, combined with the fact that updrafts and downdrafts are dynamically and microphysically (via detrained condensate) linked, support the idea that updraft size and mass flux influences downdraft size, mass flux, and by extension, downward transport of midlevel air properties that affect mesoscale convective system kinematic, thermodynamic, and microphysical evolution.

6. Conclusions

We have highlighted a pathway by which under-resolved simulated deep convective updrafts produce

FIG. 16. 2.5-km altitude mean updraft (a) mean vertical wind speed, (b) maximum vertical wind speed, (c) condensed water content, and (d) latent heating as a function of updraft horizontal area between 0400 and 0600 UTC from 34° to 35°N latitude (solid lines). The updraft area distributions for each simulation are dashed. Distributions are computed by averaging each downdraft variable (mean vertical wind speed, minimum vertical wind speed, condensed water content, and latent cooling) binned by downdraft area with bin sizes equaling the size of a single grid box (0.5626 km^2). Squares represent mean values and diamonds represent median values of all updraft samples. The 0.75-km run is shown in red and the 0.25-km run in blue.

biases in simulated mesoscale convective evolution for a leading convective, trailing stratiform squall-line case in Oklahoma during the MC3E field campaign in spring 2011. This pathway operates as follows:

- 1) Underresolved deep convective updrafts are too wide.
- 2) Wider updrafts have greater mass fluxes and carry more condensate than narrower updrafts.
- 3) Wider updrafts are associated with wider downdrafts that have greater mass fluxes and condensate than narrower downdrafts.
- 4) A greater number of relatively wide downdrafts more efficiently transport dry midlevel air downward than relatively narrow downdrafts, accelerating the development of cold pools and downward transport of horizontal momentum.
- 5) In the case of a squall line, the altered postsquall cold pool and vertical wind shear structures resulting from underresolved convective drafts interact differently with the presquall environmental vertical wind shear, affecting deep convective tilt, front-to-rear detrainment, and convective line propagation.

These impacts on MCS evolution ultimately will depend on environmental conditions, notably the wind profile. For this 20 May 2011 case, upper-level zonal winds were stronger than lower level winds, producing excessive downward transport of positive zonal momentum in the coarse-resolution run relative to the fine-resolution run and observations. This reinforced the cold pool circulation in the coarse-resolution run such that it surged forward, producing deep convection that was too upshear tilted, and midlevel front-to-rear

FIG. 17. Time series between 0400 and 0700 UTC of 2.5-km altitude convective updraft (solid) and downdraft (dashed) (a) total mass flux, (b) number, and (c) mean area. For 0400–0630 UTC, the domain considered extends from 34° to 35° N latitude west of 99° W. After 0630 UTC, the domain is extended to 98.5° W to encapsulate the squall line moving eastward. The 750-m run is shown in red and the 250-m run in blue.

detrainment that was too strong. Interestingly, kinematic, thermodynamic, and microphysical properties of convective drafts as functions of convective draft area were similar in both resolution simulations when analyzed on a common grid scale. The coarser-resolution simulation simply had a greater number of relatively large convective drafts, and larger drafts had greater mass fluxes, vertical wind speeds, condensate mass, and latent heating (for updrafts) or cooling (for downdrafts). Updraft and downdraft number, width, and total mass flux were also correlated in time, suggesting variation in downdraft size was at least partially controlled by updraft size. This result is consistent with previous studies of convective thermodynamic and dynamic behavior (e.g., Lucas et al. 1994; Coniglio et al. 2011; Marion and Trapp 2019). These results collectively imply that mesoscale convective evolution differences between the two runs in which the only difference was the horizontal grid spacing were a result of a greater number of relatively wide updrafts and downdrafts in the coarse-resolution simulation than in the fine-resolution simulation.

Some sensitivity in the results may be expected based on the definition of convective drafts, defined here using constant vertical velocity and condensate thresholds. However, we performed tests using vertical velocity thresholds of 1, 4, and 8 m s^{-1} rather than 2 m s^{-1} , and our conclusions still hold. In addition, some sensitivity can be expected to the choice of parameterizations at each grid spacing. Nonetheless, the results clearly highlight the importance of convective draft sizes and vertical transport to realistically simulate mesoscale kinematic and microphysical evolution. As large-eddy simulations (LES) of deep convection become more commonplace, they will increasingly be used to evaluate coarser-resolution runs, but this also

highlights the importance of model evaluation by observations. In the case of deep convective clouds, measurements of draft sizes along with kinematic, thermodynamic, and microphysical properties as a function of size are extremely limited. Multi-Doppler retrievals are mostly limited to case studies with insufficient resolution to quantify convective draft sizes or match instantaneous microphysics to time-integrated kinematics (e.g., Oue et al. 2019). Aircraft (e.g., Lucas et al. 1994) and radar vertical profile retrievals have much better resolution than multi-Doppler retrievals and are spread across many different cases but are limited in sample size and time–length dimensionality (e.g., Giangrande et al. 2013, 2016). Much greater sample sizes are needed to characterize updraft and downdraft spatial characteristics as a function of environmental humidity, stability, and wind properties such that LES can be validated and used more reliably to investigate additional convective cloud issues.

There are also questions related to the general applicability of these results to different deep convective cases. Are models at cloud system resolving horizontal grid spacings of 0.5–5 km biased over a range of environmental conditions toward wider, stronger downdrafts that too efficiently transport horizontal momentum and relatively low equivalent potential temperature air downward? Does this bias the development and evolution of simulated cold pool temperature, depth, and momentum over a variety of environments? Does this bias simulated MCS evolution toward particular modes such as squall lines? There is likely variability in the impacts of convective draft size biases on MCS evolution depending on ambient kinematic and thermodynamic conditions. Therefore, more cases over a range of conditions need to be analyzed

with a mix of observational retrievals, LES, and cloud system resolving models, particularly during early deep convective stages before convective draft biases affect mesoscale circulations and subsequent convective draft properties.

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