Effects of Urban Plume Aerosols on a Mesoscale Convective System

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ABSTRACT

This study examines the effects of urban aerosols on a mesoscale convective system (MCS) in the central Great Plains with the Weather Research and Forecasting Model coupled with chemistry (WRF-Chem). Urban emissions from Kansas City, Missouri, were scaled by factors of 0.5, 1.0, and 2.0 to investigate the impact of urban aerosol load on MCS propagation and strength. The first half of the storm development is characterized by a stationary front to the north of Kansas City (phase I; 1800 UTC 26 May–0600 UTC 27 May), which develops into a squall line south of the urban area (phase II; 0600–1800 UTC 27 May). During phase I, doubling urban emissions shifts the precipitation accumulation, with enhancement downwind of the storm propagation and suppression upwind. During phase II, a squall line develops in the baseline and doubled emissions scenarios but not when emissions are halved. These changes in MCS propagation and strength are a function of cold pool strength, which is determined by microphysical processes and directly influenced by aerosol load. Overall, changes in urban emissions drive changes in cloud microphysics, which trigger large-scale changes in storm morphology and precipitation patterns. These results show that urban emissions can play an important role in mesoscale weather systems.

1. Introduction

Aerosols impact Earth’s climate by reducing incoming solar radiation and altering cloud properties (Boucher et al. 2013). Depending on aerosol size and composition, these particles can act as cloud condensation nuclei (CCN) in warm clouds and/or ice nuclei (IN) in cold clouds, affecting the cloud microphysics and resulting precipitation (Stevens and Feingold 2009). Increasing aerosol concentrations can increase cloud drop number and decrease cloud drop size, which can affect cloud optical properties [the first indirect effect (Twomey 1977)]. Through decreasing cloud drop size, the number of CCN can also change precipitation patterns and cloud lifetimes, which can increase the lifetime of the cloud [the second indirect effect (Albrecht 1989)]. On weather time scales, the amount of aerosol present in a given meteorological system affects the efficiency of cloud microphysics and precipitation development, thereby altering the macrostructure and lifetime of the cloud (Stevens and Feingold 2009).

Aerosol impacts on cloud microphysics have been documented in observational studies with satellite and ground-based data (Rosenfeld 2000; Yang et al. 2011; Christensen and Stephens 2012; Min et al. 2014; Rosenfeld et al. 2014). In warm, shallow, precipitating clouds, observations and models indicate that the addition of hygroscopic aerosols suppresses precipitation (Lohmann et al. 1999; Rosenfeld 2000), as increased CCN drives competition between drops for water vapor and subsequently reduces drop sizes (Twomey 1974). In mixed-phase clouds, aerosol effects are more complex because of multiple phase changes and the associated thermodynamic processes simultaneously occurring in different locations (Seifert and Beheng 2006). The addition of CCN has competing effects on precipitation...
depending on whether warm or cold precipitation processes dominate; for example, warm precipitation processes will be suppressed, but cold precipitation processes can either be suppressed or enhanced (Cheng et al. 2010). Theoretically, precipitation enhancement has been explained by the phase change, release of latent heat, and subsequent ice processes during the transport of liquid mass to freezing levels that increases the updraft velocity (Rosenfeld et al. 2008; Koren et al. 2010; Lebo and Seinfeld 2011; Tao et al. 2012). Stevens and Feingold (2009) theorize that many coarser-resolution models may overestimate the sensitivity of cloud microphysics to aerosols because of the assumptions made in the microphysics schemes. These assumptions do not include processes that buffer the effects of aerosol perturbations, which are seen in both the observations and very-high-resolution models. For example, a CCN increase can decrease precipitation efficiency but also invigorate convection, which can increase precipitation amounts. These competing or buffering processes can be negligible or impactful, depending on the atmospheric state and synoptic conditions (Peters et al. 2014). In severe weather systems, these effects likely become more complex as aerosols can invigorate deep convection (Andreae et al. 2004; Fan et al. 2013; Storer et al. 2014) and influence intense convective precipitation events (Ntelekos et al. 2009). These nonlinear relationships between the microphysics and dynamics of aerosol–cloud interactions become more complex during strong convective events (Seifert et al. 2012).

Severe weather is defined as an event with any one of three features: 1) hail equal to or exceeding 1.0 in. (2.54 cm) in diameter, 2) wind speeds exceeding 58 m h\(^{-1}\) (26 m s\(^{-1}\)), or 3) a tornado. For each of these three cases, the atmospheric environment must be highly unstable with large amounts of convective available potential energy (CAPE) and wind shear values of 15–20 m s\(^{-1}\) over a depth of 0–6 km (Thompson et al. 2003). One type of system that often results in severe weather is the mesoscale convective system (MCS) (Maddox 1980; Houze 2004). MCSs are responsible for a large amount of growing season precipitation in the central Great Plains (CGP) (Ashley et al. 2003). They frequently form squall lines, or linear MCSs characterized by strong thunderstorms with heavy precipitation, damaging straight-line winds, hail, and occasionally tornadoes (Markowski and Richardson 2011). Idealized simulations indicate that the organization and maintenance of a squall line is a balance between the vorticity produced by cold pools and environmental wind shear (Rotunno et al. 1988). Cold pools are an important aspect of squall-line development and propagation and are defined as evaporatively cooled areas of downdraft air that spread out horizontally underneath a precipitating cloud (Engerer et al. 2008). Their strength is determined by the environment of the middle troposphere, the temperature and humidity stratification (Alfaro and Khairoutdinov 2015), and by microphysical processes within the cloud, such as sublimation, melting, and evaporation of precipitation falling through unsaturated air (Corfidi 2003) and is likely influenced by aerosol–cloud interactions (Grant and Van den Heever 2015).

Urban areas have been observed to influence local weather and climate by increasing precipitation downwind (Changnon 1981; Shepherd and Burian 2003; Jin et al. 2005; Schmid and Niyogi 2013). Observational and modeling studies have developed several mechanisms to explain this phenomenon, as reviewed in Shepherd (2005), including 1) an increase in low-level convergence and convection from the higher surface roughness of urban areas; 2) a destabilization of the boundary layer from the urban heat island (UHI) effect, which can trigger localized circulations or UHI-generated convective clouds, and 3) higher aerosol concentrations over urban areas that increase the number of CCN and lifetime of clouds.

Urban aerosol concentrations are typically higher than background aerosol concentrations because of anthropogenic activity, and aerosol–cloud interactions resulting from these emissions have the potential to alter the growth of clouds and precipitation. In one modeling study, an increase in precipitation downwind of Houston, Texas, was attributed to local meteorological feedbacks (e.g., land-use change causing an enhanced sea breeze) over increased urban aerosol concentrations (Carrió et al. 2010). However, other studies have demonstrated that urban aerosols can increase downwind precipitation through invigoration of convection (Van den Heever and Cotton 2007; Han et al. 2012). Additionally, the morphology of a storm can change as it traverses a city (Niyogi et al. 2011), which can modify when and where severe weather occurs. Ntelekos et al. (2009) attributed changes to precipitation patterns in intense convective precipitation to aerosol-induced convective invigoration. A recent study by Fan et al. (2015) found that absorbing aerosols can increase extreme precipitation events. All of these studies suggest that urban aerosols exhibit a range of impacts on precipitation and large-scale systems. Here, we focus on understanding the role of hygroscopic urban aerosols on large systems that produce severe weather, examining not only precipitation patterns but also the relationship between these aerosols, cloud microphysics, and storm dynamics.

Understanding the mechanisms that drive the interactive effects of local weather and urbanization on
severe weather is critical to predicting the societal impacts of changing urbanization and weather patterns. We explore the role of aerosols on severe storms in the urban area of Kansas City, Missouri, a metropolitan area with a population of over 2 million in 2013 (U.S. Census Bureau 2013; http://factfinder.census.gov) located on the eastern edge of the U.S. CGP. This region is known for frequent warm-season convective weather events, such as tornadoes, squall lines, hail, and heavy precipitation. Regional aerosol sources include a combination of anthropogenic sources from midsized urban areas, light industries, and energy generation embedded within an agricultural land-use matrix. With the combination of unique aerosol sources and frequent severe weather, the CGP region is well suited to investigate the urban aerosol–weather interaction. To address the question of how urban aerosols affect storm morphology, cloud microphysics, and the potential for severe weather, we scale typical urban emissions during a severe weather event and evaluate the concurrent aerosol–cloud microphysical processes and characteristics. By taking this sensitivity test approach, we reconcile the finescale differences in the precipitation response and address the influence of urban aerosols on severe weather. We describe the experiment and model simulations (section 2), analyze the precipitation and microphysical differences (section 3), and provide a discussion of our results and concluding remarks (section 4).

2. Model description

a. Model and domain

To produce a realistic evolution of the temporal formation and chemical composition of aerosols, we use the fully interactive Weather Research Forecast (WRF) Model coupled with chemistry (WRF-Chem v. 3.6; Grell et al. 2005) to simulate aerosol–cloud interactions during a severe weather event over the CGP on 27 May 2013. During this event, several reports of hail (>1.0 in. in diameter), strong winds (>60 kt), and tornadoes were recorded (NCDC 2013). A model domain with 4-km grid spacing is centered over Kansas City, Missouri (MO), and extends over an area of $988 \text{ km} \times 636 \text{ km}$ (158 rows $\times$ 246 columns; Fig. 1), with 72 vertical levels and a 50-hPa model top. We run continuous simulations from 0600 UTC 25 May through 1800 UTC 27 May with a dynamical time step of 15 s and chemistry time step of 60 s. We analyze the last 30 h of the simulation to allow for chemistry spinup in the model, based on calculations of transport time across the domain and similar studies (Eidhammer et al. 2014). For lateral meteorological boundary conditions, we use NAM-Reanalysis 12-km data (NCEI 2013; http://nomads.ncdc.noaa.gov/data.php) every six hours. Although a resolution of 250 m may be required to resolve the convective updrafts (Bryan et al. 2003), a 4-km grid size has been shown to be sufficient for resolving the potential for severe weather (Weisman et al. 1997), and we do not incorporate a cumulus scheme based on this horizontal resolution. Table 1 describes the full suite of physical and chemical parameterizations implemented in the simulation.

Simulated atmospheric aerosols are linked to microphysical processes with the Morrison microphysics scheme (Morrison et al. 2005), a two-moment bulk microphysics scheme. Aerosols are activated as CCN in a separate module and provided as condensation nuclei (CN) to the microphysics (Abdul-Razzak and Ghan 2000), which tracks the mass mixing ratio and number concentration of five hydrometeor species and water vapor (mass only): 1) cloud drop, 2) raindrop, 3) ice, 4) snow, and 5) graupel. Sources and sinks of these hydrometeors include accretion, heterogeneous freezing,
melting, self-collection, sublimation, evaporation, deposition, condensation, autoconversion (cloud drops to raindrops and ice to snow), homogeneous freezing (cloud drops to ice and raindrops to snow), ice nucleation from freezing of aerosol, and ice multiplication (riming processes) (Morrison et al. 2005). Deposition nucleation is not included, as aerosols are assumed to be coated in sulfate (Cziczo et al. 2009). Ice nucleation is not explicitly included, although this process has been shown to be important for convection (Van den Heever et al. 2006).

We include the direct and indirect radiative effects of aerosols with anthropogenic emissions and online chemistry to predict CN. In this version of the model, the chemistry module calculates the number and mass of aerosols that will activate as CCN based on hygroscopicity and supersaturation (Köhler curves), and provides this prognostic CCN to the microphysics module. Gas-phase chemistry is simulated with the Regional Acid Deposition Model, version 2, (RADM2) chemical mechanism (Stockwell et al. 1990), and particle-phase chemistry is simulated with the Modal Aerosol Dynamics Model for Europe (MADE; Schell et al. 2001), including sulfate, nitrate, ammonium, black carbon, and organic carbon. New particle formation is included via nucleation of sulfuric acid and the formation of secondary organic aerosol (SOA) is simulated with the Secondary Organic Aerosol Model (SORGAM; Schell et al. 2001).

b. Emissions and chemical boundary conditions

We implement the U.S. Environmental Protection Agency (EPA) 2011 National Emissions Inventory (NEI; EPA 2011) regridded to the 4-km model domain. The NEI includes emissions from point, area, and mobile sources for 57 gas-phase species and 19 aerosol species (including sulfate, nitrate, salt, elemental carbon, and unspcciated PM$_{2.5}$). Biogenic emissions of isoprene are estimated using the Model of Emissions of Gases and Aerosols from Nature (MEGAN) emissions algorithm (Guenther et al. 2006) and are used in the gas-phase chemistry scheme. However, the formation of SOA via isoprene is not included in this version of the model. We do not include an interactive, wind-driven source of dust [e.g., the Goddard Chemistry Aerosol Radiation and Transport model (GOCART; Zhao et al. 2010); however, the NEI includes emissions of anthropogenic dust from mobile, point, and area sources in the coarse aerosol mode. Chemical boundary conditions are provided every 6 h from the Model for Ozone and Related Chemical Tracers (MOZART-4)/GEOS-5 simulations (Emmons et al. 2010).

To investigate the sensitivity of an MCS to urban emissions and aerosols, we conduct three separate three-member ensemble simulations, each with a different emissions scenario for the Kansas City region: 1) 2011 NEI (hereinafter the BASE case), 2) doubled NEI emissions (the 2X case), and 3) halved NEI emissions (the HALF case). Because of the complexity of aerosol concentrations on cloud microphysics, especially in severe weather events, we focus on the role of urban aerosols on precipitation in a representative CGP urban environment. Emissions are scaled only in the Kansas City, MO, region (black box outlined in Fig. 2), and all NEI gas and aerosol emissions are scaled equally. Chemical boundary conditions are not altered; therefore, background aerosol and gas-phase constituents are consistent across all three simulations. Ensemble members are created by randomly perturbing the initial condition on model perturbation temperature for every grid cell. Perturbation values are created by a function that generates random numbers using a normal distribution, with a prescribed variance of 0.05 and an average of 0.0. This is slightly modified from the method provided by the WRF Working Group 13 (NOAA 2007), which is designed for operational forecasting and more ensemble members.

c. Boundary layer aerosol loading and composition

Boundary layer (surface to approximately 2 km) aerosol is averaged over the second 24-h period in the simulation (e.g., 0600 UTC 26 May–0600 UTC 27 May) for the BASE ensemble. This time slice allows time for chemical spinup yet is before the precipitation begins to accumulate. In the boundary layer, PM$_{2.5}$ concentrations are highest in the northern portion of the domain,
FIG. 2. BASE case boundary layer (0–2 km), 24-h-averaged (a) total PM 2.5, (b) sulfate aerosol, (c) ammonium aerosol, and (d) nitrate aerosol. The black box denotes the region where emissions scaling occurs in the 2X and HALF ensembles. All values are $\mu$g m$^{-3}$. 

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with values up to 50 µg m\(^{-3}\) in southern Iowa and Nebraska and up to 30 µg m\(^{-3}\) along the Missouri River valley (Fig. 2a). Much of the PM\(_{2.5}\) in the northern portion of the domain is composed of sulfate (Fig. 2b) and is driven by the interaction of the relatively large amount of ammonia emitted from the agricultural region in the northwestern part of the domain and incoming sulfate from the northeastern boundary. Evaluations of AirNow hourly station data PM\(_{2.5}\) in this region of the domain show that the model overproduces PM\(_{2.5}\) (https://www.epa.gov/hesc/remote-sensing-information-gateway). However, this pattern is consistent in all simulations. The maximum ammonium aerosol mixing ratios exceed 5 µg m\(^{-3}\) in the northwestern portion of the domain along the Missouri River valley (Fig. 2c). Distinct nitrate aerosol point sources are visible and originate in northern Oklahoma, spread out, and travel north with prevailing southerly wind into Kansas and Missouri (Fig. 2d). The spatial patterns of nitrate aerosol show well-defined plumes associated with power generation and urban areas (Fig. 1d) and comprise up to 35% of the PM\(_{2.5}\) plumes southwest of Kansas City, MO. Nitrate concentrations reach a maximum (>2 µg m\(^{-3}\)) in Omaha, NE, and just east of Kansas City with relatively little nitrate aerosol in the northern portion of the domain. Directly downwind of Kansas City, there is a local maximum nitrate aerosol concentration of 2 µg m\(^{-3}\), which is on the same order as sulfate concentrations in and near the Kansas City urban region (2–4 µg m\(^{-3}\)). This pattern reflects the shift from ammonium sulfate in the relatively rural northern portion of the domain to ammonium nitrate near the urban areas.

Modeled aerosol composition in the Kansas City metropolitan area is dominated by ammonium sulfate and ammonium nitrate. In our BASE case ensemble, elemental carbon is less than 0.15 µg m\(^{-3}\) (~1% of the PM\(_{2.5}\)) within the urban region. Studies have shown that a 20% elemental carbon fraction of PM\(_{2.5}\) is a lower limit for semidirect radiative effects to be important (Meier et al. 2012), suggesting that absorption and heating by black carbon (i.e., the semidirect effect) would be minimal. Simulated organic carbon aerosol concentrations are also relatively low. Organic aerosol is derived from primary emissions and secondary formation in the atmosphere from anthropogenic species, but the SORGAM SOA module is known to underpredict SOA formation (Ahmadov et al. 2012). SOA from biogenic volatile organic compounds (VOC) are not included in our organic aerosol estimates, as SORGAM does not calculate isoprene-derived SOA, and monoterpene emissions are not included in our MEGAN simulations. Based on the relatively low forest cover in the region (Fig. 1), the formation of local biogenically derived SOA is likely not a main driver of the organic aerosol composition.

d. Evaluation of modeled aerosol composition

Using IMPROVE data (Malm et al. 1994) from the two sites within the modeling domain [El Dorado Springs, MO, (EDS) and Lake Seguma, Iowa (LAKE); Fig. 1], we compare simulated and observed aerosol composition. Because of the precipitation timing and 4-day interval of the IMPROVE data, we compare the observed versus modeled average composition prior to the synoptic event (e.g., 22–25 May 2013, which includes two samples) in Table 2. We compare this to a similar model timeframe, averaging the simulation hours 12–30, which allows chemical spinup and is prior to the precipitation event.

At the EDS site, elemental carbon comprises 0.35 µg m\(^{-3}\) of the aerosol compared with 0.13 µg m\(^{-3}\) simulated by the model. Elemental carbon emissions from the 2011 EPA NEI and chemical boundary conditions are from anthropogenic activity only, and this discrepancy at the rural IMPROVE sites may be due to the lack of biomass-burning emissions. Evaluation of thermal anomalies in the region show that sources within the domain are small, but there are indications of biomass burning to the south in Mexico (results not shown). Because the background aerosol is consistent through all simulations and the focus of this work is on the role of urban aerosols, this discrepancy should not significantly affect our sensitivity tests near the urban area. Prior to the storm, the model and observations compare well for ammonium sulfate, with the model contributing to 2.29 µg m\(^{-3}\) of aerosol and the observed 1.43 µg m\(^{-3}\). Together, the most hygroscopic aerosols (ammonium sulfate and ammonium nitrate) contribute approximately the same fraction of the mass in both the

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**Table 2. IMPROVE aerosol and modeled aerosol (µg m\(^{-3}\)).** EDS and LAKE (Fig. 1) aerosol composition as compared to model data (20 km × 20 km averaging regions).

<table>
<thead>
<tr>
<th></th>
<th>Ammonium sulfate</th>
<th>Ammonium nitrate</th>
<th>Soil–Anthra</th>
<th>Elemental carbon</th>
</tr>
</thead>
<tbody>
<tr>
<td>EDS observed</td>
<td>1.43</td>
<td>0.29</td>
<td>0.35</td>
<td>0.35</td>
</tr>
<tr>
<td>EDS modeled</td>
<td>2.29</td>
<td>0.32</td>
<td>1.20</td>
<td>0.13</td>
</tr>
<tr>
<td>LAKE observed</td>
<td>1.24</td>
<td>0.74</td>
<td>0.19</td>
<td>0.34</td>
</tr>
<tr>
<td>LAKE modeled</td>
<td>9.82</td>
<td>1.45</td>
<td>0.53</td>
<td>0.09</td>
</tr>
</tbody>
</table>
modeled and the observed composition, which is the primary consideration for activation as CCN and is most relevant for our analysis.

In contrast, the Lake Seguma modeled data show greater discrepancies with the observations. Modeled elemental carbon is 0.09 \( \mu g \cdot m^{-3} \), while observed values are 4 times as large at 0.34 \( \mu g \cdot m^{-3} \), which we also attribute to the lack of biomass-burning emissions in our inventory and an underestimation of the Fire Inventory from NCAR (FINN), which is used to drive the MOZART model (Pereira et al. 2016). Simulated ammonium nitrate (1.45 \( \mu g \cdot m^{-3} \)) compares well with observations (0.74 \( \mu g \cdot m^{-3} \)); however, the modeled ammonium sulfate (9.81 \( \mu g \cdot m^{-3} \)) is nearly 8 times larger than the observed value (1.24 \( \mu g \cdot m^{-3} \)). This discrepancy is likely due to an interaction between local NH\(_3\) emissions from agriculture with H\(_2\)SO\(_4\) (sulfuric acid) entering from the northern boundary. One possible reason for the high ammonium sulfate formed in the model is that the MOZART model implements emissions from the 2002 EPA NEI (http://bio.egr.uiowa.edu/arctas/emission.html; Z. Lu 2016, personal communication) but not the sulfur dioxide reductions observed in the central United States (Hand et al. 2012). However, because we use the same boundary conditions for each simulation and this region is not located near the urban region of our analysis (Fig. 2), the ammonium sulfate created in the northern portion of the domain does not significantly affect our analysis.

3. Results

In the central United States, the Great Plains low-level jet consistently influences the region’s weather and climatology (Bonner 1968). The prevailing south-southwesterly wind flow transports moisture from the Gulf of Mexico to the north and has been connected with severe weather systems in the central United States. Prior to the 27 May 2013 event, the jet transports surface emissions and resulting oxidation products and aerosol from the urban regions northward. This influence is apparent in Fig. 2d, where visible plumes from the nearby urban areas extend from the point of origin (e.g., Omaha, NE, and Kansas City, MO) to the north as a result of the jet direction. Aloft, the large-scale MCS simulated in this case study originates in the northwest and propagates south to southeast. Therefore, because of these two contrasting directions, the MCS interacts with aerosols originating from the urban area well before traversing the Kansas City region. A full description of the case study follows, comparing the BASE ensemble simulation with the HALF and 2X urban emissions ensembles for the meteorology (section 3a), the accumulated precipitation (section 3b), the microphysics in phase I (0000–0600 UTC; section 3c) and phase II (0600–1800 UTC; section 3d) of the simulation, and the thermodynamics (section 3e).

a. Meteorology of the convective event: 0000–1800 UTC 27 May 2013

Observed radar reflectivity (UCAR 2015) shows strong, isolated precipitation occurring in the northeastern portion of the domain (Fig. 3a). At 0400 UTC, the precipitation in the northeastern portion of the domain is more organized, and an MCS enters the domain at the northwestern boundary (Fig. 3b). By 0600 the MCS has propagated southeast and there is an outflow boundary just north of Kansas City, MO (Fig. 3c). At 1000 UTC, the MCS continues to propagate and is now just north of Kansas City, MO (Fig. 3d). Severe thunderstorms moved across Kansas City at 1200–1400 UTC 27 May 2013 (Figs. 3e,f). The model qualitatively simulates these events, as shown in Figs. 4a–g with the BASE ensemble simulated radar reflectivity, a quantity derived from an NCAR Command Language function (wrf_user_getvar) that computes the maximum radar reflectivity over all model levels. This function uses raindrop, water vapor, graupel, and snow mixing ratios to compute the simulated radar reflectivity. This storm system divides into two distinct meteorological events: 1) a stationary front that dominates precipitation during phase I (1800 UTC 26 May–0600 UTC 27 May); and 2) an MCS that propagates through the region in phase II (0600–1800 UTC 27 May). Phase I begins with several convective cells that occur along a weak stationary front that stretches from northwestern to central Missouri at 0200 UTC (Fig. 4a). At 0400 UTC, an MCS develops in the northwest corner of the domain (Fig. 4b) and begins to grow and propagate southeast. This system merges with the frontal convection by 0500 UTC (not shown), forming an outflow boundary that travels south along the Missouri River at 0600 UTC (Fig. 4c). This time point is the end of phase I and the beginning of phase II. The outflow boundary crosses the city as it moves southwest, dissipating by 1000 UTC (Fig. 4d). Convective cells that originated with the MCS strengthen and develop into a squall line by 1200 UTC (Fig. 4e). As this squall line propagates in the southeast direction across the urban region of Kansas City, it strengthens just south of the urban area (1400 UTC; Fig. 4f). At 1500 UTC (Fig. 4g), the squall line begins to dissipate and travels into eastern Missouri, exiting the domain.

When scaling urban area emissions by factors of 2 (2X ensemble: Figs. 4h–n) and 0.5 (HALF ensemble: Figs. 4o–u), the model simulates changes in storm morphology. At 0400 UTC (46 h into the simulation), all three cases...
are qualitatively similar, with a developing MCS in the northwestern quadrant and the frontal convection along the Missouri River valley. The 2X and BASE ensemble simulated radar reflectivities are very similar through 1000 UTC. After 1000 UTC, the 2X ensemble diverges from the BASE, and the BASE and HALF simulated radar reflectivities become more similar. Noticeable differences between the BASE and HALF ensembles begin at 0600 UTC with the progression of the outflow boundary. The HALF case has lower precipitation intensity and smaller spatial extent than the BASE ensemble (Fig. 4q). From 0800 to 1000 UTC, the band of

Max Radar Reflectivity [dBz]

Fig. 3. Observed radar reflectivity (dBZ) over the central Great Plains at 0200–1500 UTC 27 May. Retrieved from the Mesoscale and Microscale Meteorology image archive (http://www2.mmm.ucar.edu/imagearchive/).

Fig. 4. Simulated radar reflectivity (dBZ) at selected time intervals over the duration of the simulation for the (a)–(g) BASE case ensemble average (h)–(n) 2X case ensemble average, and (o)–(u) HALF case ensemble average. Time intervals based on convective events discussed in the text. Phase I (1800 UTC 26 May–0600 UTC 27 May) and phase II (0600–1800 UTC 27 May) are delineated. Kansas City, MO, is shown with a black marker. The pink line is the 12 m s$^{-1}$, 10-m maximum wind speed contour.
precipitation ranging southwest to northeast of the city dissipates more quickly in the HALF ensemble (Fig. 4r). In the BASE case, there is an area of precipitation spanning Kansas City (Fig. 4d), yet this region of precipitation is smaller in the HALF case (Fig. 4r) and larger, with stronger reflectivities, in the 2X case (Fig. 4k). The strongest winds are still north of the city, with strong winds in the 2X and BASE cases extending eastward (2X farther than the BASE) and the HALF slightly farther south, covering a smaller area than the other two ensembles.

By 1200 UTC the strongest reflectivities are north of Kansas City in the BASE case, with a developing system extending southwest of the city. While the system north of the city in the HALF case is stronger, it lacks the southwestern development seen in both the BASE and 2X cases (Fig. 4s). The strongest reflectivities are shifted to the south in the 2X ensemble (Fig. 4l), indicating a stronger squall-line development southwest of Kansas City. This is confirmed by the 12 m s\(^{-1}\) contour, where there is a distinct pattern, with all three ensembles producing strong winds to the north of the city. However, the 2X and BASE ensembles also include strong winds extending from the city center to the southwest, and the HALF ensemble fails to produce this feature (Figs. 4e,l,s). This pattern persists through 1400 UTC, where the 2X and BASE ensembles have strong winds southwest of Kansas City, indicating a possible squall line (Figs. 4f,m). This is not present in the HALF ensemble (Fig. 4t) and is consistent with the simulated radar reflectivity. By 1500 UTC, the first pulse of convective activity has dissipated and moved south, and the main squall line crosses the Kansas City, MO, region in all three cases (Figs. 4g,n,u).

b. Accumulated precipitation

The total accumulated precipitation pattern can be attributed to two meteorological events: 1) the stationary front that dominates precipitation during phase I and 2) the MCS that propagates through the region in phase II. In the following discussion, we refer to the direction of storm propagation relative to Kansas City; upwind is to the northwest of the city, and downwind is to the southeast. Phase I accumulated precipitation is clearly associated with the stationary front (Figs. 5a–c), with maximum accumulation exceeding 80 mm in the 12-h period (Fig. 5a). The swath of precipitation extends northwest to southeast, with the leading edge of precipitation reaching northeastern Kansas City. In the 2X ensemble, accumulated precipitation is reduced up to 50 mm [80\%] (Fig. 5b). Differences in the HALF ensemble exhibit greater spatial variability with no clear pattern associated with the emissions decrease (Fig. 5c).

For the final 12 h of the simulation (phase II; 0600–1800 UTC; Figs. 5d–f), we focus on the region of precipitation exceeding 65 mm near the Kansas City urban area. This region of precipitation extends east and south over Kansas City and is associated with a squall-line development (Fig. 5d). In the 2X case, there is suppression east of the city (a decrease of 40 mm, or 50\%) and enhancement (20 mm, or >100\%) to the north and south of the urban area (Fig. 5e). In the HALF case, there is a distinct suppression of precipitation over Kansas City extending east and south, with values exceeding 30 mm (50\%). North of Kansas City precipitation is enhanced up to 20 mm (Fig. 5f). Comparison with the storm morphology suggests that changes in the spatial distribution of precipitation are due to a displacement of the squall line, which occurs north of Kansas City as compared to the BASE case.

c. Spatial distribution and time series of hydrometeors: Phase I

By varying the Kansas City anthropogenic emissions, we directly affect aerosol concentrations and therefore the available CCN within and upwind of the city. As described in sections 3a and 3b, this magnitude change leads to large-scale changes in storm morphology and accumulated precipitation patterns. To understand how scaling urban emissions affects the spatial distribution of hydrometeors on a cloud macroscale, we examine the model microphysics via changes in liquid water path (LWP; kg m\(^{-2}\)) and ice water path (IWP; kg m\(^{-2}\)). LWP and IWP are each temporally integrated over all hours in phase I.

During phase I, LWP and IWP are elevated along the stationary front that extends from northwest to southeast just to the north of Kansas City (Fig. 6a). The 2X ensemble LWP decreases to the northeast and increases to the southwest of the front as compared to the BASE case (Fig. 6b). This result is consistent with warm rain suppression theory, where more aerosols initially increase the cloud drop number, decreasing the cloud drop size and slowing the autoconversion to rain. The displacement of the system to the south suggests that it takes longer for raindrops to form, and when they do, they drive enhancement downwind compared to the BASE case. Solid-phase hydrometeors form via heterogeneous or homogenous ice nucleation. The Morrison microphysics scheme employed considers heterogeneous ice nucleation to occur by condensation freezing of deliquesced, internally mixed particles or homogenous ice nucleation, which is expressed as a probability (Morrison et al. 2005). Homogenous ice nucleation occurs with temperatures colder than −40°C; however, ice nucleation onto aerosol is not included in the model. As a result of solid-phase
formation in the microphysics parameterization, the change in IWP is similar to the LWP. In the HALF case, no clear spatial pattern is evident in either LWP (Fig. 6c) or IWP (Fig. 6f), although the magnitude of change is comparable to the 2X case. The broader spatial impacts in the IWP versus the LWP are a result of the surface changes propagating to higher altitudes and being dispersed by stronger winds aloft.

Next, we compare the temporal changes in the hydrometeors during phase I. In Fig. 7, we average over all
model vertical levels and over the region shown in Fig. 5a that encompasses the region where the stationary front occurred and a majority of the accumulated precipitation, while excluding grid cells with less than 1 drop per cubic centimeter and the elevated PM$_{2.5}$ in the northern portion of the domain (Fig. 2a). At 0000 UTC 27 May, all three ensembles have a similar accumulated precipitation (~15 mm). The HALF ensemble initially accumulates precipitation more quickly than the BASE or 2X cases. By the end of phase 1, the BASE ensemble
has accumulated 32 mm of precipitation, the HALF ensemble has accumulated 30 mm of precipitation, and the 2X ensemble has accumulated 29 mm of precipitation. The BASE ensemble is usually between the 2X and HALF ensembles, with the minimum/maximum shading overlapping, suggesting that these changes may not be significant. The cloud drop number concentration and size are also sensitive to the emissions scaling. The 2X ensemble has the largest concentration (152 cm$^{-3}$), the BASE ensemble has a concentration of 135 cm$^{-3}$, the HALF ensemble starts with 130 cm$^{-3}$ (Fig. 7b), and the ensemble shading for the 2X case is outside of that for the BASE and HALF cases. This pattern of 2X, BASE, and HALF persists until 0000 UTC 27 May, at which point, the HALF case cloud drop number concentrations decrease more rapidly than the 2X or BASE ensembles. As phase I progresses, cloud drop sizes grow larger and number concentrations decrease to ~70 cm$^{-3}$.

Droplet size is increasing as number concentrations are decreasing for all ensembles. Raindrop number concentrations and sizes increase for all three ensembles throughout phase I (Fig. 7c); raindrop size increases from 100 to 430–440 μm. For the first half of phase I, the 2X ensemble generally has fewer but slightly larger raindrops than the HALF or BASE ensembles; after the halfway mark, the 2X raindrops are more numerous and continue to be larger, although this does not exceed the range of variability of the other two ensembles.

Graupel number concentrations remain relatively stable throughout most of phase I and decrease to about 13 L$^{-1}$ by the end of phase I (Fig. 7d). Initially graupel sizes differ between the ensembles [2X (~400 μm) > BASE and HALF (~50 μm)], but all ensembles are about 450 μm by 2200 UTC 26 May and remain steady through the end of phase I. All three ensembles increase

Fig. 7. Phase I time series of spatially and vertically integrated hydrometeor number concentrations (solid lines; left y axes) and time series for the spatially and vertically averaged hydrometeor effective radius (dashed lines; right y axes; μm) for the BASE (black), 2X (red), and HALF (blue) cases. (a) Accumulated precipitation (mm). (b) Cloud drop number concentration (cm$^{-3}$) and size. (c) Raindrop number concentration (L$^{-1}$) and size. (d) Graupel number concentration (L$^{-1}$) and size. (e) Snow number concentration (L$^{-1}$) and size. (f) Ice number concentration (cm$^{-3}$) and size. (g) Accumulation mode interstitial aerosol number concentration (cm$^{-3}$). Shading indicates the minimum/maximum of the ensemble members.
in snow number concentrations and size until 2200 UTC 26 May (Fig. 7e). The differences in number concentrations and sizes are relatively small, especially after 0000 UTC 27 May. Finally, ice hydrometeor number concentrations increase through 0000 UTC 27 May for all three ensembles (Fig. 7f), and ice hydrometeor size increases throughout phase I. As with the snow number concentrations and sizes, differences are especially small between the ensembles after 0000 UTC 27 May.

To demonstrate how scaling emissions affects the aerosols and subsequently the available CCN, we show the interstitial accumulation mode aerosol number concentrations (Fig. 7g). Total accumulation mode aerosol number concentration would include the in-cloud category as well; however, those aerosols are passed into the microphysics as activated CCN (i.e., cloud drops). At the beginning of phase I, all three ensembles have similar aerosol number concentrations (∼190 cm⁻³). As precipitation forms and the sources continually emit, by the end of phase I, there are more aerosols in the 2X ensemble than in the BASE or HALF ensembles, although this is not outside of the model spread.

To summarize phase I, scaling emissions leads to changes in the liquid-phase hydrometeors consistent with current warm rain theory; increasing emissions leads to increasing the available CCN, which leads to smaller cloud drops and delayed precipitation or suppression upwind and enhancement downwind of the stationary front. For the HALF ensemble, the spatial changes are less coherent, though the time series indicate that fewer activated CCN lead to larger initial cloud drops and earlier precipitation. However, the changes in the solid-phase hydrometeors are relatively small and do not scale directly with emissions. The rates at which these hydrometeors form and grow set the stage for phase II.

d. Spatial distribution and time series of hydrometeors: Phase II

During phase II (0600–1800 UTC 27 May), a squall line develops south of Kansas City in the BASE and 2X ensembles and is weaker in the HALF ensemble. This phenomenon was shown in the simulated radar reflectivity (Fig. 4) and the accumulated precipitation differences (Fig. 5). Here we link these changes in storm morphology and precipitation to the spatial and temporal hydrometeor characteristics during phase II.

In the BASE ensemble, the temporally integrated LWP and IWPs of 30 and 85 kg m⁻², respectively, are associated with the squall line and are largest south of Kansas City (Figs. 8a,d). The 2X ensemble LWP and IWP values show suppression of the squall line upwind and enhancement of the squall line downwind of Kansas City (Figs. 8b,e), and the HALF ensemble displays a strong reduction of the LWP and the IWP [<20 kg m⁻² (~66%) and <45 kg m⁻²] south of Kansas City (Figs. 8c,f). The corresponding increase to the north of the city is not as large. The large decrease in IWP in the HALF ensemble is due to the suppressed squall-line development.

As in phase I, we compare the temporal changes in the hydrometeor size and number concentrations. In Fig. 9, we average over all model vertical levels and over the region shown in Fig. 5d, which encompasses the region where the squall line occurred and excludes grid cells with less than 1 drop per cubic centimeter. At the beginning of phase II (0600 UTC), accumulated precipitation is similar for all three ensembles (~13.5 mm; Fig. 9a). By the end of phase II, there is a clear enhancement of mean accumulated precipitation in the 2X ensemble outside of the HALF and BASE ensemble spread and a reduction of precipitation in the HALF ensemble. At 1000 UTC, cloud drop number concentrations are the same for all three ensembles (70 cm⁻³), and the 2X drop size is clearly larger than the BASE and HALF ensembles (15 vs 13 μm), which generally persists throughout phase II. After 1100 UTC, cloud drop number concentrations decrease for all three ensembles, as cloud drops are converted to rain or solid-phase species (Figs. 9b,c,d,e,f). As a result of less precipitation, the HALF ensemble has more cloud drops than both the 2X and BASE ensembles from 1100 through 1400 UTC, with little overlap with the other ensembles. Raindrop number concentrations are increasing (Fig. 9c), coinciding with the squall-line development (0600–1200 UTC). The raindrop number concentrations are similar until 1200 UTC, increasing to 130 L⁻¹ (BASE and 2X) and 150 L⁻¹ (HALF); then the BASE and 2X ensembles continue to increase until 1400 UTC. The 2X raindrop sizes are consistently larger than the BASE and HALF ensembles. Greater vertical velocities (not shown) in the 2X ensembles can support larger hydrometeors and suggest a more efficient collision and coalescence process. Graupel number concentrations increase through 1100 UTC, with the 2X ensemble once again being more numerous than the HALF and BASE ensembles during most of that time. After 1100 UTC, 2X graupel number concentration decreases and is distinct from the other ensembles, but the size increases through 1400 UTC. Number concentrations of snow and ice are steady throughout phase II, with 2X number concentrations and sizes of snow and ice generally greater than both BASE and HALF snow and ice number concentrations and sizes, although there is little distinction between the ensembles. The urban emissions scaling is most evident in the 2X ensemble aerosol number concentrations at
0600 UTC (~160 vs ~125 cm\(^{-3}\); Fig. 9g). As precipitation increases in the 2X ensemble and ceases in the BASE and HALF ensembles (Fig. 9a), aerosol number concentrations increase in the BASE and HALF ensembles and remain steady in the 2X ensemble until 1000 UTC. After the squall line passes at 1400 UTC, 2X ensemble aerosol number concentrations increase much more quickly than in the BASE and HALF ensembles. While meteorology is a strong driver of the diurnal cycle of aerosol number concentrations, they are also influenced by the diurnal pattern of emissions, as seen in all three ensembles.

During phase II, increasing urban emissions leads to convective invigoration [i.e., enhanced precipitation, hydrometeor growth, and number concentrations (Fig. 5e)], and decreasing emissions leads to the suppression of the squall line (Fig. 5f). The 2X ensemble has the largest hydrometeors, while the HALF and BASE ensembles are similar. From 1000 to 1400 UTC, the HALF ensemble has more numerous and smaller cloud drops, coinciding with fewer and smaller solid-phase hydrometeors (Figs. 9b,c,d,e).

In summary, the growth rates of hydrometeors and subsequent timing of the synoptic meteorology drive differences in the simulated squall-line development. At the beginning of phase II, more numerous and smaller raindrops contribute to the development of the squall line differently in the HALF ensemble compared to the 2X or BASE ensembles. This difference leads to more substantial changes in the growth and formation of solid-phase
hydrometeors in the 2X and BASE ensembles, which feeds back into the formation of rain from cold cloud processes. The link between hydrometeors and the weakened squall-line development lies in the connection between microphysics and cold pools, as squall-line formation and development is a function of the strength of cold pools and their propagation speeds. We next examine how the rate of hydrometeor evaporation determines cold pool strength.

e. Dynamics: Cold pools

Cold pool strength determines the propagation speed of a squall line, which controls its development and subsequent propagation. The strength of a cold pool is determined by the thermodynamic profile of the atmospheric column; the difference between the equivalent potential temperature near the surface and at the origin of the downdraft (Schlemmer and Hohenegger 2014). Because the equivalent potential temperature accounts for the condensation of all water vapor, changes in cold pool strength can be related to microphysical processes. Here we use the perturbation temperature \( T' \) as a proxy for cold pool strength (Kalina et al. 2014). We calculate \( T' \) as the difference between the lowest model level perturbation temperature and the domain-averaged lowest-model-level perturbation temperature, where a larger negative value of \( T' \) corresponds to a stronger cold pool. We use \( T' \) and the 10-m maximum wind speed to examine the relationships between the cold pool strength and squall-line formation.

Here we show that cold pool magnitude and expanse (location) are sensitive to urban emissions. At 1000 UTC, a weak cold pool (\( \approx 2 \) K) northwest of Kansas City develops in the BASE ensemble (Fig. 10a); this cold pool is much stronger in the 2X ensemble (\( \approx 6 \) K; Fig. 10e) and is stronger (\( \approx 3 \) K) but displaced south in the HALF ensemble (Fig. 10i). The 12 m s\(^{-1}\) contour follows this pattern, encompassing the largest area with the strongest cold pool in the 2X ensemble. By 1100 UTC, \( T' \) decreases to \( \approx 4 \) and to \( \approx 8 \) K for the BASE and 2X ensembles, respectively (Figs. 10c,g). While the strength of the HALF ensemble cold pool is comparable to the BASE ensemble, it continues to be displaced to the south and is ahead of precipitation (Figs. 10k, 5d). This pattern is reflected in

**Fig. 9.** As in Fig. 7, but for phase II (0600–1800 UTC 27 May).
the 12 m s\(^{-1}\) contour, which is to the north of Kansas City in the BASE ensemble but extends southward in the HALF ensemble. By 1200 UTC, the squall line has developed south of Kansas City (Fig. 3f), approximately located along the \(23^\circ\) contour (Fig. 10d) and along the 12 m s\(^{-1}\) contour. A similar pattern can be seen in the 2X ensemble, except the cold pool is slightly stronger (\(-5\) K contour; Fig. 10h). In the HALF ensemble, the cold pool is stronger (\(-4\) K) and much farther southeast (Fig. 10k). As a result of this southern displacement of the cold pool, the squall line is greatly weakened, leading to reduced amounts of precipitation in the HALF ensemble. Therefore, changes in cold pool strength and location between ensembles are likely a function of the changes in the diabatic processes driven primarily by changes in the microphysics and aerosol concentrations.

4. Discussion and conclusions

This study examined the effects the magnitude of urban emissions have on a severe weather system in the CGP, a region with midsized urban areas and subject to a large number of severe weather events. Emissions within the Kansas City urban area (Fig. 2) were doubled (2X ensemble) and halved (HALF ensemble) from a baseline (BASE) emissions scenario. We examined a specific severe weather event (26–27 May 2013) and organized the analysis in two phases, referring to a stationary front in the first half of the simulation (phase I) and the formation and propagation of an MCS in the second half of the simulation (phase II).

The perturbation to urban emissions directly affects the aerosol concentrations throughout the domain, as the low-level jet in the region transports direct aerosol emissions as well as those formed through atmospheric processes from the urban area to the north. The resulting effects of these emissions changes are described here and visually in Fig. 11. Changes in aerosol concentrations affect the number of CN that activate as CCN, which affects the hydrometeor formation and growth rate. Initially, during phase I, increased aerosol concentrations (2X ensemble) shift the total accumulated precipitation to the southwest, where rainfall is enhanced downwind and suppressed upwind. Later, in
Fig. 11. Schematic that describes the processes affected by changing urban emissions. Aerosol concentrations are directly affected by scaling urban emissions, which lead to changes in the number of CCN available and the growth and formation of hydrometeors. The growth and formation of hydrometeors affect the dynamics of the squall-line changes. The differences in the squall-line propagation ultimately lead to a different precipitation spatial pattern and altered aerosol concentrations.

In our simulations, we attempt to understand the meteorology associated with realistic atmospheric chemistry simulations that include dynamic aerosol number concentrations, composition, and evolution of aerosol burden. Our three-member ensemble approach (e.g., three simulations for each of the three emissions scenarios) lends strength to the consistent response of the meteorology to the change in aerosol magnitude. By showing the variance between the ensemble members, we gain confidence in asserting how the changes between each emissions scenario lead to changes in the meteorology. Many of the prior studies aimed at untangling the thermodynamic and dynamic processes involved in cloud microphysics and cloud macrostructure are idealized simulations (Morrison and Grabowski 2013; Grant and Van den Heever 2015). While idealized simulations provide insight into how a single process might affect a convective system, they are limited in their applications, as they often do not include surface heterogeneity and in many cases include a constant CCN concentration (activated vs nonactivated) that is prescribed (Van den Heever and Cotton 2007; Carrió et al. 2010). By using the WRF-Chem, we provide a direct link to dynamic aerosol concentrations that drive cloud microphysics and close the feedback loop shown in Fig. 11 that allows the precipitation to alter aerosol concentrations themselves.

We do note several caveats in our simulations. The model does not include ice nucleation for any aerosol type, which may affect the formation of solid-phase hydrometeors and the cold pool analysis. Additionally, we do not account for dust emitted from the surface naturally, which would potentially have a large impact, as dust is an important ice-forming nucleus. Because we are investigating severe weather, and several studies show evidence for convective invigoration, it would be useful for future studies to include direct aerosol IN, especially in locations with potentially large amounts of dust, such as the CGP. The nucleation of ice via dust might have significant radiative impacts and change the cloud microphysics. Additionally, as noted in section 2d, there are lower-than-observed black carbon concentrations within our domain at the two sites with speciated aerosol observations. This could be a result of the omission of biomass-burning emissions within the domain or the underestimation of emissions within the MOZART simulations that drive the boundary conditions. These aerosols also may have important implications for the thermodynamic (semidirect) response, and prior studies have shown that severe weather can be impacted by the presence of black carbon (Saide et al. 2015). While the analysis presented in this paper is focused on the second indirect effect, we recognize that the first direct effect is potentially impactful. The changes to the anvils as a result of scaling emissions would affect the radiation budget, influencing the stability of the atmosphere. These effects could be especially important in longer-term simulations (Fan et al. 2013). Finally, we note that we include chemical boundary conditions in these simulations to provide as much realism in the aerosol evolution as possible. However, our evaluation shows that we are likely simulating too much sulfate at the northern boundary of the model domain, triggering higher-than-observed concentrations of ammonium sulfate. Our three emissions scenarios would be impacted equally by the emissions and boundary conditions above, and we focus on processes in and around the urban area that are removed from the driving boundary conditions. Specifically, we have identified how the urban emissions at the domain
center can affect the propagation of the large-scale meteorology.

Overall, we find that changing the magnitude of urban emissions has the potential to alter severe weather and rain patterns. During phase I, increasing the urban emissions leads to a spatial shift in the precipitation pattern, while decreasing urban emissions leads to comparatively little change. The shift in total precipitation in the 2X ensemble can be attributed to slowed warm rain processes. Conversely, during phase II, the altered urban emissions have the opposite effect: decreasing urban emissions suppresses the squall-line formation. We attribute the suppression to changes in cold pool strength and earlier precipitation, variables that are affected by microphysical processes and thus aerosol concentrations. Considering that the precipitation response to scaling emissions (and aerosol concentrations) is dependent on event, this type of analysis could be performed on several severe weather events, the results of which could be used to determine a threshold for how increasing or decreasing urban aerosols can influence regional precipitation. These changes in the precipitation patterns and locations of severe weather could impact urban planning and agriculture, as this study suggests that severe weather can be sensitive to the magnitude of urban emissions. 

The use of realistic, evolving aerosol concentrations within these simulations shows the effects that the concentrations have on different stages of the storm and indicates the nonlinearity of responses during the storm lifetime.

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REFERENCES


