The dynamical effects of increased aerosol loading on the strength and structure of numerically simulated squall lines are explored. Results are explained in the context of Rotunno–Klemp–Weisman (RKW) theory. Changes in aerosol loading lead to changes in raindrop size and number that ultimately affect the strength of the cold pool via changes in evaporation. Thus, the balance between cold pool and low-level wind shear–induced vorticities can be changed by an aerosol perturbation. Simulations covering a wide range of low-level wind shears are performed to study the sensitivity to aerosols in different environments and provide more general conclusions. Simulations with relatively weak low-level environmental wind shear (0.0024 m s$^{-1}$) have a relatively strong cold pool circulation compared to the environmental shear. An increase in aerosol loading leads to a weakening of the cold pool and, hence, a more optimal balance between the cold pool– and environmental shear–induced circulations according to RKW theory. Consequently, there is an increase in the convective mass flux of nearly 20% in polluted conditions relative to pristine. This strengthening coincides with more upright convective updrafts and a significant increase (nearly 20%) in cumulative precipitation. An increase in aerosol loading in a strong wind shear environment (0.0064 m s$^{-1}$) leads to less optimal storms and a suppression of the convective mass flux and precipitation. This occurs because the cold pool circulation is weak relative to the environmental shear when the shear is strong, and further weakening of the cold pool with high aerosol loading leads to an even less optimal storm structure (i.e., convective updrafts begin to tilt downshear).

1. Introduction

Recently, the sensitivity of deep convective clouds to anthropogenic aerosol perturbations has received considerable attention in the literature (e.g., Khain et al. 2004; Khain and Pokrovsky 2004; Khain et al. 2005; Wang 2005; Koren et al. 2005; Grabowski 2006; Seifert and Beheng 2006; Teller and Levin 2006; Van den Heever et al. 2006; Fan et al. 2007; Tao et al. 2007; Van den Heever and Cotton 2007; Khain et al. 2008; Lee et al. 2008b,a; Rosenfeld et al. 2008; Fan et al. 2009; Khain and Lynn 2009; Koren et al. 2010; Noppel et al. 2010; Ekman et al. 2011; Lee 2011; Lebo and Seinfeld 2011; Grabowski and Morrison 2011; Seifert et al. 2012; Morrison 2012; Tao et al. 2012; Lebo et al. 2012; Storer and van den Heever 2013). Because aerosols are fundamentally linked to the formation of cloud droplets and ice crystals in our atmosphere, it should come as no surprise that changes in the ambient concentration of these particles may lead to changes in cloud properties. However, because of the complexity of these systems (i.e., mixed-phase microphysics, dynamical feedbacks, radiative feedbacks, etc.), understanding how an increase in ambient aerosol number concentration affects macroscale features is highly challenging. Our understanding of the effects of increased aerosol loading on the warm region of these clouds is rooted in the work of Gunn and Phillips (1957) and Squires (1958) whereby it was shown that increases in aerosol number suppress collision–coalescence and thus mitigate the formation of precipitation. Yet, more recent studies have suggested that even in warm clouds the effects of aerosols can be quite complex due to feedbacks between microphysics and dynamics (e.g., Ackerman et al. 2004; Lu and Seinfeld 2005; Wood 2007; Bretherton et al. 2007; Chen et al. 2011). Moreover, the complexities of the extensive mixed-phase
region in deep convective clouds result in numerous other changes that may ultimately lead to a net increase or decrease in precipitation.

Several studies have suggested that deep convection is intensified in polluted compared to pristine environments (e.g., Tao et al. 2007; Fan et al. 2009; Rosenfeld et al. 2008; Lebo and Seinfeld 2011; Lebo et al. 2012), while others indicated little to no sensitivity (e.g., Khain and Lynn 2009; Morrison 2012). Convective invigoration in these studies has often been explained through increased latent heating in polluted conditions (e.g., Van den Heever et al. 2006; Rosenfeld et al. 2008; Khain and Lynn 2009; Lebo and Seinfeld 2011; Lebo et al. 2012). Mechanistically, this is caused by a reduction of droplet collision–coalescence, leading to lofting of liquid water above the freezing level that in turns drives enhanced freezing and ice processes (e.g., Rosenfeld et al. 2008).

More recently, Lebo et al. (2012) described the enhancement of latent heating in polluted relative to pristine conditions because of larger condensation rates directly associated with higher droplet concentrations and smaller mean droplet size. Other studies have related the intensification or weakening of convection to changes in cold pools and low-level convergence (e.g., Tao et al. 2007; Lee et al. 2008b; Seigel et al. 2013). Significant disagreement among modeling studies of aerosol effects on deep convection likely reflects in part the underlying complexity of these systems, with numerous interacting microphysical and dynamical processes leading to complementary or competing effects (Morrison 2012).

The overall sensitivity of the strength of deep convection to aerosols is also likely related to the environmental conditions and dynamical characteristics of the system being analyzed. For example, Tao et al. (2007) examined tropical convection using a two-dimensional (2D) cloud resolving model (CRM) with bin microphysics and showed that precipitation is enhanced in moist environments and suppressed in drier environments; the authors related these changes to differences in rain evaporation rates and cold pool strength. However, Lebo and Seinfeld (2011) found no such sensitivity to relative humidity for supercells using a three-dimensional (3D) CRM with bin microphysics. Fan et al. (2009) demonstrated using a 2D CRM with bin microphysics that in environments with weak environmental vertical wind shear, an increase in aerosol loading acts to enhance convection while in strong-shear environments, the same increase in aerosol loading has little effect or even suppresses convection. More generally, environmental shear exerts a dominant control on storm type, with increasing shear favoring single-cell, multicell, and supercell storms (see, e.g., Markowski and Richardson 2010).

Of particular interest is the sensitivity of mesoscale convective systems (MCSs), specifically squall lines, to aerosol loading. Squall lines are a type of linear MCS that commonly occurs in the tropics and midlatitudes. They are responsible for producing heavy precipitation, large hail, damaging straight-line winds, and occasional tornadoes. A key aspect of squall-line organization and maintenance is the inherent balance between the cold pool strength and environmental wind shear (e.g., Thorpe et al. 1982; Nicholls et al. 1988; Weisman et al. 1988; Rotunno et al. 1988; Fovel and Ogura 1989; Szeto and Cho 1994; Robe and Emanuel 2001; Weisman and Rotunno 2004; James et al. 2005; Bryan et al. 2006; Takemi 2007). Rotunno et al. [1988, hereafter referred to as Rotunno–Klemp–Weisman (RKW) theory] discuss in detail how the optimal state of a squall line exists in an environment where the contribution of vorticity from the cold pool balances the contribution of vorticity from the low-level environmental shear. If the shear is too weak, the line will tilt in the upshear direction. On the other hand, if the cold pool is too weak, the squall line tilts in the downshear direction. This is important because upshear and downshear tilting can lead to weakening of updrafts from enhanced entrainment of dry environmental air and adverse perturbation pressure gradient forces (Markowski and Richardson 2010; Parker 2010). Changes in cold pool strength via changes in microphysical processes ought to lead to changes in the low-level dynamics and hence the strength and organization of squall lines (Fig. 1).

The parameterization of rain microphysics in particular has been noted as an important factor in the strength and maintenance of squall lines given the impact of rain evaporation on cold pool characteristics (e.g., Ferrier et al. 1995; Morrison et al. 2009; Bryan and Morrison 2012; van Weverburg et al. 2012). For example, Morrison et al. (2012) found that the strength and speed of a squall line was sensitive to the raindrop breakup parameterization implemented in a bulk microphysics model. Given that the ambient aerosol number concentration leads to substantial changes in droplet number and thus collection processes, it is postulated that increased anthropogenic aerosol loading could have similar effects on squall-line dynamics by impacting cold pool evolution. Since aerosols can affect cold pool characteristics because of their impact on cloud microphysics, RKW theory provides a potentially useful conceptual framework by which to analyze aerosol effects on squall lines. These effects are schematically summarized in Fig. 1. Figure 1 serves two purposes: 1) it provides a conceptual framework of a typical squall line and 2) it suggests pathways for aerosols to affect squall-line dynamics. The proposed effects are confirmed and
described in detail in sections 4 and 5. Briefly, however, the key idea is that aerosols ultimately affect the raindrop size distribution which in turn alters the bulk rain evaporation rate and cold pool intensity. These effects alter the balance of the cold pool–induced circulations with the low-level environmental shear to produce an intensification (weakening) of the squall line in relatively weak (strong) wind shear environments.

Before proceeding, it is important to keep in mind the potential shortcomings of RKW theory as discussed in Stensrud et al. (2005) and Bryan et al. (2006). In particular, Stensrud et al. (2005) noted that observed long-lived, severe squall lines were often far from the optimal state. They also pointed out that some measures of system strength (e.g., total and maximum vertical velocity) in the simulations of Rotunno et al. (1988) and Weisman and Rotunno (2004), which served as a basis for RKW theory, did not peak near the optimal state. Other issues include the role of mid- to upper-level shear (e.g., Fovell and Ogura 1995; Parker and Johnson 2004) as well as the applicability of RKW theory to broader environments because the simulations of Rotunno et al. (1988) and Weisman and Rotunno (2004) used only a single thermodynamic sounding. Moreover, several studies have simulated long-lived squall lines in suboptimal states (e.g., Fovell and Ogura 1988, 1989; Lafore and Moncrieff 1989; Rotunno et al. 1990; Coniglio and Stensrud 2001; Weisman and Rotunno 2004), despite the original applicability of RKW theory in Rotunno et al. (1988) to explain the longevity of squall lines. Nonetheless, simulations from more recent squall-line studies have generally supported RKW theory (Bryan et al. 2006; Morrison et al. 2012). However, Morrison et al. (2012) found that while structure and intensity of convective updrafts were consistent with RKW theory, surface precipitation peaked in suboptimal conditions. A thorough and systematic study that addresses all of these potential issues is beyond the scope of the present work.

While the sensitivity of aerosol effects on deep convective systems to environmental shear has been
previously described (e.g., Lee et al. 2008b; Fan et al. 2009), these studies did not provide a dynamical explanation for this sensitivity. Thus, the broad motivation for our study is to provide a dynamical context for investigating aerosol effects on squall lines, especially in terms of sensitivity to environmental shear. The specific goals of this study are to

1) quantify the enhancement or suppression of convection and precipitation in a squall line due to increased aerosol number concentration;
2) provide a conceptual framework for the dynamical effects of increased aerosol loading on squall lines by exploring sensitivity of these effects to environmental wind shear in the context of RKW theory.

To address these points, we focus our attention on the changes in cold pool strength produced by increasing the aerosol number concentration and how it relates to the environmental shear. Before presenting the simulated results, section 2 provides background information on the microphysics model, domain setup, and chosen squall-line case; section 3 introduces key aspects of RKW theory necessary for analyzing the model simulations performed here. Two particular cases, one representing a weak shear scenario and the other a strong shear scenario are discussed both conceptually and quantitatively in section 4. Section 5 is reserved for the analysis of both changes in precipitation and convective strength within the aerosol number concentration–low-level environmental shear parameter space. Last, we present the important conclusions from this work in section 6.

2. Methods

a. Dynamical framework

The bulk microphysics model of Morrison et al. (2009) [as adapted by Lebo et al. (2012), see below for more details] is coupled to the Weather Research and Forecasting Model (WRF), version 3.3.1 (Skamarock et al. 2008), for use as a 3D “cloud resolving” model. The model is compressible and nonhydrostatic. The model domain is defined to be 124 km × 714 km in the meridional and zonal directions, respectively. The domain is extended to 20 km in the vertical. Horizontal grid spacing is 1 km. There are 80 levels in the vertical; the vertical grid spacing is about 0.25 km. To maintain dynamical stability, the time step is set at 2.5 s for all simulations with a duration of 8 h. Rayleigh damping is applied to the uppermost 5 km of the domain. The boundaries are set to be open in the zonal direction and periodic in the meridional direction. Since the simulations are specifically designed to be idealized, we neglect the effects of radiation, surface fluxes, and Coriolis acceleration in the present study. Advection of scalars is calculated using a fifth- and third-order monotonic advection scheme in the horizontal and vertical, respectively.

b. Microphysics model

The simulation of aerosol impacts on clouds of any type relies on several key components contained within the cloud microphysics model. Here, the assumption that all points within a model domain are at saturation if condensed water is present (the saturation adjustment assumption) is relaxed so that the supersaturation is predicted prognostically (analogous to the method used in more detailed bin microphysics models). Without the prediction of supersaturation, it was shown in Lebo et al. (2012) that the bulk model did not qualitatively agree on the sign of the change in the convective mass flux and precipitation compared to bin model simulations for an increase in aerosol loading. However, the inclusion of a prognostic supersaturation algorithm in a bulk microphysics scheme (Morrison et al. 2009) provided good quantitative agreement with detailed bin model simulations for the changes in convective mass flux with aerosol loading in a supercell. Given this, and that high-resolution bin microphysics simulations over a large domain and longer time scales are computationally prohibitive (especially if one wants to perform numerous simulations for various environmental parameters), we restrict the simulations here to include bulk microphysics, namely the scheme of Morrison et al. (2009) as modified in Lebo et al. (2012) to include prognostic supersaturation.

The bulk microphysics scheme is also modified to explicitly represent aerosol activation and regeneration [see Lebo and Seinfeld (2011) for more details]. Briefly, the aerosol model represents the ambient aerosol distribution via 36 mass-doubling bins with a smallest aerosol diameter of 0.01 μm. Activation is predicted by applying Köhler theory to the unactivated aerosol population. To determine the size of the newly activated droplets over the course of a single time step, we apply the assumption of Kogan (1991) as used by Khain et al. (2000) and Xue et al. (2010) in which the newly formed droplet size is simply a factor k larger than the dry aerosol diameter. Bulk mass and number mixing ratio tendencies due to activation are then added to the bulk tendencies from other cloud processes.

Xue et al. (2010) demonstrated the importance of aerosol regeneration on cloud properties in orographic clouds. Here, aerosol regeneration refers to the process by which a cloud droplet completely evaporates within a time step and thus reproduces an aerosol particle. Following Mitra et al. (1992), it is assumed that the evaporation
of a single cloud droplet results in the regeneration of a single aerosol particle. The regenerated aerosols are assumed to be distributed identical to the initial aerosol distribution. This assumption implies that the effect of collection processes is small on the evolution of the aerosol distribution. More sophisticated methods for incorporating the effects of cloud processes on aerosols are currently being developed (e.g., Lebo and Morrison 2013). However, exploration of such effects is beyond the scope of the present study.

c. Model setup

Convection is triggered in the model by applying forcing directly to the vertical velocity $w$ field over the first hour of the simulations (Ziegler et al. 2010). The forcing is applied within a half cylinder with radii of 10 km in the $x$ direction and 2.5 km in the $z$ direction; a maximum acceleration of 0.5 m s$^{-1}$ is located at the center of the half cylinder. (Note that the $w$ forcing is uniform in the line-parallel dimension.) The $w$ forcing decays radially from the center assuming a cosine function of the radius. Random thermal perturbations (amplitude 0.1 K) to the initial sounding are applied within a region 40 km wide in the $x$ direction centered around the region of $w$ forcing and 4 km deep to initiate 3D motion. This is unlike previous works in which convection is initiated via the application of a warm or cold bubble in the lower troposphere. We choose to initiate convection in this manner so as to allow the low-level dynamics to spin up unimpeded by initial warm or cold bubbles that could potentially lead to changes in the quasi-steady-state cold pool. To allow the squall line to spin up, we restrict most of the analysis to the final 4 h of the simulations (i.e., between 4 and 8 h, unless stated otherwise). Since all simulations are spun up for 4 h and use the same forcing parameters, the effects of this assumption on the overall results should be minimal.

The CRM is initialized with the sounding (Fig. 2) from the observationally based squall-line case study of the Eighth International Cloud Modeling Workshop held in July 2012 in Warsaw, Poland (Muhlbauer et al. 2013). The initial sounding comprises the 0000 UTC 20 June 2007 soundings from Lamont, Oklahoma (LMN), below 700 hPa and from Norman, Oklahoma (OUN), above 700 hPa to represent the prestorm environment. The sounding is smoothed using a 1–2–1 smoother with 20 iterations. The chosen sounding has a convective available potential energy (CAPE) of about 6800 J kg$^{-1}$ based on the most energetic parcel. The observed $\Delta u$ from wind-profiler data in Purcell, Oklahoma, was between 12 and 14 m s$^{-1}$ over a depth of 5 km. Details of the observed squall line from this case are given by Morrison et al. (2012).
WRF simulations using the Morrison et al. (2009) microphysics scheme with observations for this case using a similar model setup. In particular, the model is able to reproduce the overall observed reflectivity field with a well-defined convective core and trailing stratiform region. Further comparison of the simulations with observations is beyond the scope of this study; a more detailed comparison of several model simulations with observations for this case is being performed as part of the World Meteorological Organization (WMO) Cloud Modeling Workshop (Mühlbauer et al. 2013).

To test the sensitivity of squall lines to aerosol perturbations, simulations are performed with initial aerosol number concentrations of 100, 200, 500, 1000, and 2000 cm⁻³, encompassing relatively clean maritime-influenced air to rather polluted urban-influenced air. Aerosols are assumed to be soluble ammonium bisulfate. The sensitivity to aerosol composition is beyond the scope of the present study. The aerosols are assumed to be lognormally distributed with a standard deviation and geometric mean diameter of 1.8 and 0.08 μm, respectively. Most of the analysis below revolves around a comparison of cases in which Na₅⁻¹⁰⁰ or 1000 cm⁻³ (clean and polluted scenarios, respectively). The parameter space is further explored in section 5.

d. Sensitivity simulations

We performed an additional set of simulations in which saturation adjustment was employed for calculations of cloud water condensation and evaporation. These simulations help to isolate effects of increased latent heating aloft from changes in the cold pool by neglecting the dependence of the droplet condensation rate (and the associated latent heating) on the mean cloud droplet size, thereby limiting the mechanism of increased latent heating in polluted conditions. Previous work has shown that the response of the convective mass flux to aerosol perturbations using the Morrison et al. (2009) bulk microphysics scheme with saturation adjustment is either negligible or slightly negative for supercells (e.g., Lebo et al. 2012; Morrison 2012). Thus, if the primary effect of aerosol perturbations on squall lines is to affect the low-level dynamics, the simulations using saturation adjustment ought to exhibit enhanced convection as well. On the other hand, if the primary effect of an increase in aerosol loading is to alter the latent heating rates aloft and feedback on the dynamics of the system, the sensitivity simulations with saturation adjustment ought to show little sensitivity to aerosol number concentration. These simulations are performed only for the relative weak and strong wind shear environments (i.e., Δu = 12 and 32 m s⁻¹).

A final set of sensitivity simulations were performed using soundings with less CAPE (down to 2000 J kg⁻¹ based on the most energetic parcel) to confirm that the results presented below can be generalized for squall lines (not shown). The sensitivity soundings were formed by systematically decreasing the water vapor mixing ratio in the lowest 5 km to achieve a reduction in CAPE.

e. Statistical significance

Statistical significance is determined by using all points N in the domain and in time (i.e., N = nₓ × nᵧ × nz × no, where nₓ, nᵧ, nz, and no correspond to the number of points in the x, y, and z directions, and output times, respectively). Student’s t tests are performed on the means of two samples of size N corresponding to a clean and polluted scenario to rule out the null hypothesis (in other words, we are confident that the two means are not identical).

3. Cold pool–shear interaction

It is important to first discuss the dynamical context by which squall lines form and persist to carefully analyze the effects of aerosol perturbations on their strength and structure. To do so, we revisit the hallmark work of Rotunno et al. (1988) in which a theory (RKW theory) for long-lived squall lines was formulated. RKW theory suggests the importance of the balance between the contribution of vorticity from the cold pool and the contribution of vorticity from the environmental shear to squall-line structure and maintenance. To look at this relationship in more detail, the cold pool intensity c² is defined as

\[c^2 = -2 \int_0^H -B_L dz,\]  \hspace{1cm} (1)

where H is the top of the cold pool and B_L corresponds to the buoyancy at some point L behind the gust front. Equation (1) is derived for an idealized hydrostatic density current (c is the density current propagation speed) and thus in more realistic 3D nonhydrostatic simulations, determining the exact location of L is quite challenging. Following RKW theory and Markowski and Richardson (2010), L should correspond to the point behind the gust front in which the zonal flow relative to the gust front is stationary. Moreover, the pressure within the cold pool at L should be nearly hydrostatic. For the purpose of this present work, we choose L to be a fixed location 30 km behind the gust front and zonally average over a 20-km band around L as well as average in the meridional (line parallel) direction to get a characteristic value of c for the squall
We find (as will be shown later) that the results are qualitatively insensitive to the chosen $L$ as long as $L$ is not within 10–15 km of the gust front where large non-hydrostatic effects occur.

RKW theory suggests that a key factor in determining the structure and strength of a squall line is the ratio $c/Du$. For $c/Du > 1$ (or what will be referred to as a suboptimal shear environment), the squall line tends to lean back over the cold pool (upshear). The other extreme is the case in which $c/Du < 1$ (or what will be referred to as a suboptimal squall line). In this scenario, the line tilts downshear. The optimal state occurs when $c/Du \approx 1$. As $c/Du$ approaches unity, the line tends to be more upright and thus entrainment of the environmental air and adverse pressure perturbation effects are reduced, aiding to increase updraft velocities within the convective region.

The structure of a squall line in terms of environmental shear and $c$ is the focus of RKW theory. Here, we look at the sensitivity of squall-line structure and strength to changes in aerosol number concentration. Thus, we vary $c$ directly while varying $c$ indirectly through aerosol perturbations. For example, if $c/Du \gg 1$ and an increase in aerosol loading leads to a significant decrease in $c$, the squall line is more optimal in the polluted environment for a given $Du$ and, hence, convective updrafts should be more upright and stronger according to RKW theory. Performing this comparison for the range of environmental shear profiles described above allows us to analyze the sensitivity of a squall line to aerosol perturbations in the context of RKW theory.

4. Results

For brevity, we initially restrict the analysis to a weak shear case and a strong shear case, characterized by $Du = 12$ and $32$ m s$^{-1}$, respectively. By doing so, we can specifically address aerosol sensitivity in these distinctly different environments before exploring the entire parameter space in detail. Moreover, much of the analysis is restricted to comparing simply the 100 and 1000 cm$^{-3}$ cases for clarity, hereafter referred to as clean and polluted conditions, respectively.

a. Weak shear environment

In Fig. 3 we show profiles of spatially averaged and temporally averaged (4–7 h) convective updraft mass flux and convective updraft fraction. Here, we define the mean convective updraft mass flux as a function of height $[MF(z)]$ to be the product of the vertical velocity $w$ and the air density $\rho$ for all locations in which $w \geq 2$ m s$^{-1}$, divided by the total area of the domain. Qualitatively, the results presented here are not sensitive to the chosen threshold value for $w$. Here we see both a consistent increase in the convective mass flux and a decrease in the convective updraft fraction (defined as the fraction of the domain at a given vertical level containing updrafts of at least 2 m s$^{-1}$) for an increase in aerosol loading. As described in the introduction, invigoration of deep convection has often been explained in previous studies via the suppression of collision–coalescence and enhancement of mixed- and cold-phase processes and hence latent heating aloft in polluted compared to pristine conditions (e.g., Khain et al. 2004; Khain et al. 2004).
Rosenfeld et al. 2008; Lebo and Seinfeld 2011), or because of increased latent heating due to larger condensation rates associated with smaller mean droplet size (Lebo et al. 2012). As described later in this section, the latter mechanism contributes to the invigoration of convection in polluted conditions in the simulations here. However, compared to some previous studies of aerosol impacts on deep convection (e.g., Van den Heever et al. 2006; Lebo and Seinfeld 2011; Lebo et al. 2012; Morrison 2012), the increase in convective updraft mass flux shown here is much larger, upward of 20% at some levels. This suggests an additional mechanism leading to enhanced convection beyond just the direct change in latent heating following the mechanisms described above. This additional enhancement is caused by changes in cold pool strength with increased aerosol loading and interaction of the cold pool with environmental shear. In particular, the decrease in the convective updraft fraction with increased aerosol loading is suggestive of a narrower (at least in the horizontal plane) updraft. Morrison et al. (2012) showed a similar response to changes in the parameterization of drop breakup and related this finding to more upright updrafts consistent with RKW theory. All else being equal, if the convective updraft core is cylindrical, then as updrafts become more upright the horizontal width (at a given vertical level) approaches the radial width of the convective core. Thus, smaller horizontal widths suggest more upright updrafts. This is confirmed in Figs. 4a,b where the occurrence of more upright (and stronger) updrafts is seen with increased aerosol loading via a shift forward in the maximum of the vertically averaged convective mass flux. The shift forward is approximately 2 km. The shift is most noticeable in the midlevel maximum updraft velocity and the connection between the low-level and mid- to upper-level updrafts that is present in the polluted scenario (Fig. 4b, dark blue shading) and not present in the clean case (Fig. 4a).

To understand why the convective mass flux increases and the convective updraft fraction decreases, we turn to profiles of the condensed mass concentrations (Fig. 5a). Here we see that with increased aerosol loading (solid to
dashed) there is an enhancement of supercooled liquid water (green) and a subsequent increase in ice mass concentration aloft (black). However, Fig. 5b suggests that despite these increases in mass concentrations there are smaller cloud droplets and ice crystals on average (mean volume radius, due to the large increase in number concentration). There is very little change in the snow mass concentration and snow size (blue) but a rather substantial change in graupel (orange). Here we see that the graupel mass concentration decreases but the mean size increases (about 20% throughout most of the column). As these larger graupel particles fall, they melt and this results in larger mean raindrop size as well (red). Furthermore, since the addition of aerosol particles acts to mitigate the collision–coalescence process, the formation of numerous small raindrops via warm processes (i.e., autoconversion and accretion) is hindered. This reduces the number concentration of raindrops and helps promote the growth of larger raindrops via cold-rain mechanisms (riming and subsequent melting in particular). Thus, even though the mass evaporation rate of a single drop is proportional to its radius (in other words, a larger raindrop will evaporate mass faster than a smaller counterpart), there are far fewer raindrops on average in polluted compared to pristine conditions via cold-rain mechanisms and they fall faster leading to shorter residence time and a smaller mass concentration (although this is somewhat compensated by ventilation effects). These results (fewer larger raindrops in polluted conditions versus many smaller raindrops in more pristine air) suggests that the bulk rain evaporation rate ought to be reduced as a result of increasing the aerosol number concentration. The changes observed in the model results are shown schematically in Fig. 1 (top) where the sizes of the particles are depicted by changes in the corresponding symbol size.

Figure 6b indeed demonstrates that the increase in raindrop mean size in conjunction with a decrease in bulk mass concentration in polluted compared to pristine conditions leads to a substantial decrease in the rain evaporation rate within the cold pool. All else being equal, this leads to less negative buoyancy (Fig. 6a). Figure 6c shows that the reduction in negative buoyancy occurs throughout most of the cold pool (at least for points more than 20 km behind the surface gust front, defined to be the line-averaged mean position of the location in the x direction in which the wind direction changes sign). Moreover, Fig. 6d shows that the cold pool is slightly shallower in the polluted case (for points more than 10 km behind the gust front).

Figures 6c,d show that the region chosen to determine the value of \( c \) [namely, point \( L \) in Eq. (1)] does not qualitatively affect the change in \( c \) for an increase in aerosol loading as long it is not within 10–20 km of the gust front. Picking a location close to the gust front would have a large effect on the results, but doing so would be inconsistent with the derivation of \( c \) from density current theory as explained in section 3. We compute \( c \) over the region between 20 and 40 km behind the gust front, namely, where the cold pool top is relatively flat (Fig. 6d). Increasing aerosol loading leads to a consistent decrease in \( c/\Delta u \) and thus a more
optimal state according to RKW theory for the weak wind shear case (Fig. 6c). This decrease in $c/\Delta u$ is quite large, ranging from 10% to 20% and thus providing a mechanism for the statistically significant invigoration ($p = 7.025 \times 10^{-8}$) seen in Fig. 3a (in addition to increased latent heating caused by changes in condensation rate described below). The polluted, stronger squall line also ought to move slower (since $c$ is smaller).
compared to the clean, weaker squall line. Figure 6f confirms this hypothesis by showing the mean change of the gust front position as a function of time. Here, a negative value means that the gust front in the polluted scenario trails the gust front in the clean case. The negative slope from 5.5 h and onward shows that the polluted squall line moves slower by about 2 km h\(^{-1}\) over that time frame.

Evidence for the invigoration effects of increased aerosol loading via both latent heating increases and a more optimal balance between the cold pool intensity and low-level environmental wind shear is provided in Figs. 7 and 8. As described in section 2c, additional simulations were performed using a version of the bulk microphysics scheme with saturation adjustment. Previous work has shown that the response of the convective mass flux to aerosol perturbations using the Morrison et al. (2009) bulk microphysics scheme with saturation adjustment is either negligible or slightly negative for supercell storms (e.g., Lebo et al. 2012;}

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**Fig. 7.** Profiles of horizontally and temporally averaged (for \(4 \leq t \leq 8\) h) latent heating rates for simulations performed with (a) explicit supersaturation treatment and (b) saturation adjustment. Shown are data for the clean \((N_a = 100\, \text{cm}^{-3}, \text{solid})\) and polluted \((N_a = 1000\, \text{cm}^{-3}, \text{dashed})\) scenarios for those simulations with weak wind shear (i.e., \(\Delta u = 12\, \text{m s}^{-1}\)). The net heating (black) is shown and the heating rates are separated into warming (red) and cooling (blue) for clarity.

**Fig. 8.** (a) Time series of line-averaged \(c/\Delta u\) (as in Fig. 6e) and (b) relative change (polluted minus clean) in mean convective updraft mass flux for \(\Delta u = 12\, \text{m s}^{-1}\), except for simulations performed with saturation adjustment. Data are averaged between 4 and 7 h.
Thus, by using saturation adjustment, the effect of aerosol perturbations on latent heating through modification of mean droplet size and hence condensation rate is neglected and, thus, we can focus on the low-level cold pool effects. Figures 7a,b corroborate the results of Lebo et al. (2012). Simulations using the baseline model configuration with explicit condensation/evaporation show a substantial increase in mean latent heating in polluted compared to pristine conditions, but the simulations using saturation adjustment show little difference. Although differences in mean latent heating between polluted and pristine conditions are small in simulations with saturation adjustment, there is still a small increase in the convective mass flux (for $z > 7$ km), consistent with changes in cold pool intensity leading to a more optimal squall line according to RKW theory (Fig. 8). Specifically, the decrease in $c/\Delta u$ shown in Fig. 6e is also present in simulations performed with saturation adjustment (i.e., negligible latent heating effects) shown in Fig. 8a. Thus, it is clear that the effects of an increase in aerosol number concentration on both latent heating (because of increased condensation from smaller mean droplet size) and cold pools is to enhance convection for relatively weak wind shear conditions.

b. Strong shear environment

As the low-level wind shear increases, the ratio of $c$ to $\Delta u$ decreases for constant $c$. If the wind shear is high enough, $c/\Delta u$ can become less than 1 and, hence, less optimal. For the weak wind shear simulations described previously, $c$ was found to lie between 25 and $33 \text{ m s}^{-1}$. Thus, we choose the high-shear case to be one in which $\Delta u = 32 \text{ m s}^{-1}$ so that $c/\Delta u$ is less than unity during most of the simulation. Note that changes in $c$ do occur between the simulations. However, the change in $c$ due to a change in low-level wind shear is much smaller than the change in $\Delta u$.

Comparing Figs. 3 and 9, we see that as $\Delta u$ increases, there is a reduction in the invigoration of the convective mass flux with an increase in aerosol loading. In the high shear scenario, there is a small decrease in the convective mass flux between 2 and 5 km, with little change seen elsewhere. Furthermore, Fig. 9b shows an increase in the convective updraft fraction for an increase in aerosol number. This is opposite of the response for the weak wind shear environment described above. Because $c/\Delta u$ is less than 1 for the clean case, this means that the line should tilt more forward (see Figs. 4c,d; cf. the weak low-level wind shear scenario) under more polluted conditions. Evidence of this is again seen in Figs. 4c,d, which shows a slight forward shift of the updraft cores. According to RKW theory, this ought to result in a reduction in the strength of the line, but Fig. 9a suggests otherwise. This contradictory finding is a direct result of the increase in droplet condensation and hence latent heating due to enhanced aerosol number concentration discussed earlier that acts to compensate the weakening caused by the forward titling of the squall line (as demonstrated with the sensitivity simulations using saturation adjustment). As was also shown previously in the weak wind shear scenario, the low-level updrafts tend to be more connected with the mid- to upper-level updrafts in the polluted case (dark red contours in Fig. 4c,d).

In Fig. 10 we see that the evaporation rate in the cold pool decreases slightly with increased aerosol loading, primarily below 2.5 km. The average buoyancy in the

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Fig. 9. As in Fig. 3, but for simulations with high wind shear (i.e., $\Delta u = 32 \text{ m s}^{-1}$).
cold pool is somewhat less negative. Figures 10c,d show that the buoyancy is slightly less negative for all points behind the gust front while the cold pool depth is more or less unchanged up to 50 km behind the gust front. These small changes combine to decrease \( \frac{c}{\Delta u} \) by \(<10\%\) on average (Fig. 10e). Since \( \frac{c}{\Delta u} < 1 \) for the clean case, the reduction in \( c \) means that the system is tilted somewhat more downshear and the effects of aerosol loading on the cold pool act to weaken convection. Again, however, one has to consider all aerosol effects on the dynamics and since the buoyancy aloft increases due to enhanced latent heating (Fig. 11), the overall effect is only a slight (although statistically significant, \( p = 1.006 \times 10^{-5} \)) suppression in the vertically integrated convective mass flux (Fig. 9a). These changes are portrayed schematically in Fig. 1 (bottom).
5. General results within the aerosol number concentration–environmental wind shear parameter space

While the previous section provides evidence for the enhancement of convection in weak wind shear environments and negligible impact or a weakening of convection in strong shear environments, it is important to understand how the aerosol sensitivity varies over a range of low-level environmental wind shears. Moreover, it is important to also understand how these changes in dynamics lead to changes in precipitation. In Fig. 12 we show the relative change in convective mass flux and precipitation as a function of environmental wind shear (here again defined as the difference in the line-normal wind between the surface and the top of the shear layer at 5 km). The relative changes are shown for an increase in aerosol number from 100 cm\(^{-3}\) to the corresponding values on the y axis. We find that as the low-level wind shear decreases, squall lines are more susceptible to aerosol effects (via both enhanced convection and increased precipitation). In these cases, the line tends to tilt back over the cold pool and thus any weakening in the cold pool will bring the updraft core closer to the gust front and thus enhance convection consistent with RKW theory (see Fig. 4). We see that in the lowest shear scenario, the convective mass flux increases by at least 10% and the precipitation increases by nearly 20% for some aerosol loadings. Both of these changes are found to be statistically significant at the 0.05 significance level using a standard Student’s t test for the two simulations.

As the shear increases to between 15 and 25 m s\(^{-1}\), the simulations shift to a different regime in which the change in the convective mass flux with increased aerosol loading may be slightly positive or slightly negative (Fig. 12a). For a given shear over this range of shear values, Fig. 12b shows an increase in precipitation by 3%–5%. The reason for these opposing effects on precipitation and convective mass flux is that aloft, enhanced buoyancy due to increased latent heating in the polluted environment leads to a slight increase in the strength of convection and an increase in the total condensed mass concentration. This is offset by a weakening caused by the decreased cold pool strength (since, in the clean environment, the system is already nearly optimal and so weakening of the cold pool produces a suboptimal environment).
Thus, as a whole, the change in the convective mass flux is small, but the enhancement in mixed-phase and cold microphysical processes aloft leads to small increases in precipitation.

As the shear approaches the upper bound of the parameter space explored here, the convective mass flux decreases somewhat for large changes in aerosol loading (again, a balance between an increase aloft and decrease at low-levels—see Fig. 9a), but the precipitation strongly decreases with increased aerosol loading (Fig. 12). Changes in the convective mass flux for the highest wind shear scenario are similar to that for the middle regime discussed above, but the change in precipitation has the opposite sign. This is explained by examining Fig. 13. Here we have plotted normalized rain rates, total condensed mass concentration, and relative humidity as a function of line-normal distance from the leading edge of the gust front (thin dashed line). These figures illustrate two important points. First, as the environmental shear increases (black to blue to red), the maximum rain rate and maximum total condensed mass concentration shift closer to the gust front. Quantitatively this difference from $\Delta u = 12$ to 32 m s$^{-1}$ is about 28 and 11 km for the rain rate and total condensed mass concentration, respectively. Moreover, these results are consistent with changes in tilting of the convective updrafts with increased aerosol loading discussed previously, especially for the weak wind shear scenario (i.e., $\Delta u = 12$ m s$^{-1}$) due to the shift forward (downshear) in the maximum rain rate of between 4 and 5 km for polluted compared to pristine conditions (Fig. 13a). Second, Fig. 13a shows that as the shear increases from 24 to 32 m s$^{-1}$, the rain-rate maximum shifts toward the gust front by about 10 km, but the leading edge of the precipitation does not move as far downshear. However, Fig. 13b demonstrates that more condensed water mass exists downshear of the gust front as the shear increases (in fact, the shift forward in the maximum is about the same as the shift forward in the leading edge of the precipitation). This is important because the background environment is quite dry relative to the region within 50 km behind the gust front (Fig. 13c). In other words, more precipitation falls at or ahead of the gust front in the strongest shear case and subsequently evaporates before reaching the surface. This explains the large decrease in precipitation with increased aerosol loading for the strongest shear case. This is also consistent with the small increase of 1%–2% in the relative humidity ahead of the gust front with increasing low-level shear (i.e., the environment is moistened as a result of the downshear shift of falling precipitation and its subsequent evaporation; Fig. 13).

Figure 12 suggests that the response of squall lines to an aerosol perturbation may not be monotonic. Specific
6. Conclusions

The sensitivity of numerically simulated squall lines to aerosol perturbations is investigated across a wide range of background environmental shear profiles in the context of RKW theory. A two-moment bulk microphysics scheme (Morrison et al. 2009) with explicit treatment of supersaturation (Lebo et al. 2012) and coupled to a binned aerosol scheme (Lebo and Seinfeld 2011) was used in WRF for this study. Recent work concerning aerosol effects on deep convection has provided a mix of results (see section 1). Here we presented evidence that the effects of aerosol perturbations on deep convection are a distinct function of the environment and dynamical characteristics of the system being investigated. Thus, results presented herein are applicable to squall lines only and not necessarily other types of deep convection (e.g., while the convective cores of squall lines are strongly influenced by cold pool–induced circulations in the context of low-level environmental shear, other systems like supercells are likely to exhibit different sensitivities owing to their different dominant dynamical driving mechanisms).

The key conclusions of this study are as follows:

1) **Weak shear**—In relatively weak low-level shear environments, an increase in aerosol number concentration led to an increase in convective mass flux (strengthening), corroborating the work of Fan et al. (2009). This was caused by a weakening of the cold pool in polluted conditions since the rain mass concentration was smaller and the raindrops were larger and thus less readily evaporated. In the context of RKW theory, this weakening of the cold pool reduced the ratio of \( c/\Delta u \) toward a more optimal value consistent with stronger, more upright convective updrafts. Increased latent heating resulting from greater condensation because of the reduced mean cloud droplet size in polluted conditions also contributed to invigoration of the squall line.

2) **Strong shear**—In relatively strong low-level shear environments, an increase in aerosol loading led to fairly small changes in the convective mass flux although there was a small suppression at the highest shear values. In the context of RKW theory this is explained by a decrease of \( c/\Delta u \) from a more optimal state in pristine conditions \((c/\Delta u \sim 1)\) to a less optimal state \((c/\Delta u < 1)\) because of the reduced cold pool strength in polluted conditions. This led to greater forward (downshear) tilting of convective updrafts with increased aerosol loading and hence weakening of the updrafts. However, this weakening via cold pool–shear interactions was compensated by invigoration due to greater condensation and latent heating in polluted conditions. In the highest shear conditions and for relative large aerosol perturbations, the cold pool effects outweighed the increase in latent heating so that there was an overall suppression of the convective mass flux in polluted compared to pristine conditions. For smaller aerosol perturbations and in somewhat less strongly sheared conditions, these two effects compensated and the overall impact of aerosols on the convective mass flux was very small.

While the convective mass flux was found to be enhanced in weak shear and slightly suppressed in strong shear, the effect on precipitation is more complicated. For the weakest wind shear, it was shown that an increase in aerosol loading led to an increase in the convective mass flux because of both increased latent heating and more upright updrafts that enhanced the formation of more precipitation (condensed mass aloft increased with increasing aerosol number concentration). As the low-level wind shear was increased, the simulated squall line entered a middle regime whereby aerosols had limited effects on the convective mass flux but precipitation increased with an increase in aerosol number. The increase in precipitation resulted from an increase in latent heating due to greater condensation that ultimately led to an increase in condensed mass in polluted conditions. For the highest wind shear scenario, a third regime was evident in which increased aerosol loading led to a small decrease in the convective mass flux but a substantial decrease in precipitation. This was due to updrafts that tilted more downshear, producing precipitating hydrometeors that fell ahead of the gust front and readily evaporated.

These results provide a context by which aerosol perturbations can dynamically alter a midlatitude squall line. While results were presented in detail for a single thermodynamic profile in this study, an additional suite of simulations (not shown) with lower CAPE (down to 2000 J kg\(^{-1}\)) were performed to corroborate and
generalize the results across a wide range of conditions typically observed in midlatitude squall-line environments. An important aspect of this work is that the terms “weak wind shear” and “strong wind shear” are relative to the cold pool intensity and thus depend on the environmental sounding. Therefore, simulations with reduced CAPE tend to have less precipitation, weaker cold pools, and an optimal state that occurs at relatively weaker wind shear according to RKW theory. The robustness of the results lies in the overarching effect of an increase in aerosol loading that leads to a suppression in cold pool intensity across the low-level shear parameter space explored. Changing the thermodynamic state of the system will in fact alter the optimal point in the low-level wind shear–cold pool intensity parameter space according to RKW theory, but has limited impact on the overall aerosol effects. That said, the exploration of aerosol effects in drastically different thermodynamic environments (i.e., the tropics) is important, but is beyond the scope of the present study and should be addressed in future work.

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