The Influence of Lifting Condensation Level on Low-Level Outflow and Rotation in Simulated Supercell Thunderstorms

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ABSTRACT

This paper reports on results of idealized numerical simulations testing the influence of low-level humidity, and thus lifting condensation level (LCL), on the morphology and evolution of low-level rotation in supercell thunderstorms. Previous studies have shown that the LCL can influence outflow buoyancy, which can in turn affect generation and stretching of near-surface vertical vorticity. A less explored hypothesis is tested: that the LCL affects the relative positioning of near-surface circulation and the overlying mesocyclone, thus influencing the dynamic lifting and intensification of near-surface vertical vorticity. To test this hypothesis, a set of three base-state thermodynamic profiles with varying LCLs are implemented and compared over a variety of low-level wind profiles. The thermodynamic properties of the simulations are sensitive to variations in the LCL, with higher LCLs contributing to more negatively buoyant cold pools. These outflow characteristics allow for a more forward propagation of near-surface circulation relative to the midlevel mesocyclone. When the mid- and low-level mesocyclones become aligned with appreciable near-surface circulation, favorable dynamic updraft forcing is able to stretch and intensify this rotation. The strength of the vertical vorticity generated ultimately depends on other interrelated factors, including the amount of near-surface circulation generated within the cold pool and the buoyancy of storm outflow. However, these simulations suggest that mesocyclone alignment with near-surface circulation is modulated by the ambient LCL, and is a necessary condition for the strengthening of near-surface vertical vorticity. This alignment is also sensitive to the low-level wind profile, meaning that the LCL most favorable for the formation of intense vorticity may change based on ambient low-level shear properties.

1. Introduction

Supercell thunderstorms pose a unique scientific challenge due to their tangible socioeconomic impacts paired with the physical complexity clouding our understanding of their structure and evolution. Parameters such as convective available potential energy (CAPE) and bulk deep-layer (0–6 km) vertical wind shear are effective predictors of whether deep moist convection will be supercellular in nature (e.g., Johns et al. 1993; Thompson and Edwards 2000; Thompson et al. 2003; Craven and Brooks 2004). However, the existence of a supercell does not guarantee tornadogenesis—rather, only a fraction of supercells produce tornadoes (Trapp et al. 2005). Thus, identifying factors that differentiate nontornadic and tornadic supercells is crucial, both for our physical understanding of these storms and for forecasting and warning of tornadoes.

Two parameters in the near-storm environment have shown skill in distinguishing between nontornadic and tornadic supercells: low-level (0–1 km) vertical wind shear and lifting condensation level (LCL). Climatologies utilizing proximity soundings of supercell thunderstorms have identified increased low-level shear values in the environments of tornadic supercells, as compared to nontornadic supercells (Rasmussen and Blanchard 1998; Thompson et al. 2003; Craven and Brooks 2004). The same climatologies have also noted a tendency toward higher boundary layer relative humidity (RH) and consequently lower LCLs in tornadic supercell environments. However, the exact dynamic and thermodynamic reasons for the predictive skill of low-level shear and LCL have yet to be shown conclusively.

Previous work has suggested a few potential dynamical links between LCL and supercell tornadogenesis.
For instance, the ability of outflow parcels to be lifted and stretched is partially dependent on the buoyancy of air near the updraft (Markowski et al. 2002; Markowski and Richardson 2014). Numerous studies, both numerical and observational, have noted that significantly tornadic storms often have warmer (less negatively buoyant) rear-flank downdraft (RFD) outflow, likely due to its decreased resistance to being dynamically lifted (Markowski et al. 2002; Shabbott and Markowski 2006; Grzych et al. 2007). Consequently, the existence of colder and more negatively buoyant cold pools can have a detrimental effect on tornadogenesis. Though potentially associated with more baroclinic generation of horizontal vorticity, this colder outflow inhibits the stretching of near-surface air and can potentially undercut low-level circulations, thus reducing the effectiveness of the vertical perturbation pressure gradient force (VPPGF; Brooks et al. 1994; Gilmore and Wicker 1998; Markowski and Richardson 2014). Herein lies the predication with storm outflow; outflow needs to be negatively buoyant enough to generate baroclinic vorticity along the forward flank gust front, but not so much so that said outflow cannot be dynamically lifted (Markowski and Richardson 2009). Some recent papers (e.g., Schenkman et al. 2014; Roberts et al. 2016) have noted that the inclusion of surface drag in simulations can lead to significant frictionally generated horizontal vorticity both with and without the presence of a cold pool. However, the general importance of frictionally generated horizontal vorticity to the development and maintenance of near-ground vertical vorticity, especially relative to baroclinically generated horizontal vorticity, is still largely unknown. Furthermore, despite their occurrence in simulations, it remains unclear if mesocyclone tornadoes often form before mature cold pools have had the chance to develop in nature (Markowski 2016).

Outflow propagation—though dependent on numerous variables ranging from the storm scale down to microphysical scales—has been linked to environmental RH (Brooks et al. 1994; Adlerman and Droegemeier 2002; van den Heever and Cotton 2004; Snook and Xue 2008) and thus LCL height. McCaul and Cohen (2002) further demonstrated this relationship in an idealized environment, showing that lower LCLs (i.e., higher boundary layer RH) produce storms that are less outflow dominated. These findings are consistent with our conceptual understanding of LCL, where lower LCLs lead to decreased evaporative cooling as precipitation falls below cloud base, and therefore weakened outflow. Microphysical studies have come to similar conclusions, suggesting that reduced RFD evaporation in the presence of low LCLs leads to more buoyant outflow and increased likelihood of tornadogenesis (e.g., Kumjian 2011; French et al. 2015). Conversely, higher LCLs lead to a deeper and drier subcloud layer, thus increasing evaporative cooling and resulting in a more outflow-dominated storm (Markowski et al. 2002). This being said, some studies have suggested that increased CAPE and/or precipitable water values often associated with lower LCLs might alter precipitation distribution in a way that actually enhances evaporative cooling and strengthens outflow (Lerach and Cotton 2012).

As mentioned earlier, tornadic supercells are often characterized by increased shear within the lowest kilometer of the atmosphere. The impact of low-level shear extends beyond its magnitude, however, as both the depth and orientation of low-level shear have been shown to influence near-surface vertical vorticity in simulations (Wicker 1996; Markowski et al. 2003; Thompson et al. 2003; Guarriello et al. 2018, hereafter G18) and observations (Miller 2006; Esterheld and Giuliano 2008). Specifically, hodographs containing predominately streamwise environmental horizontal vorticity and southerly shear in near-surface layers as shallow as 500 m have been shown to favor tornadogenesis (Miller 2006; Esterheld and Giuliano 2008; Kis and Straka 2010; Nowotarski and Jensen 2013), particularly when compared with wind profiles containing a large crosswise component and easterly shear near the surface (Coffer and Parker 2017). Two hypotheses have been proposed to explain the physical link between these shear characteristics and tornadoes. First, the orientation of horizontal vorticity within a storm’s inflow, as determined by the mean vertical shear, influences the formation of the low-level mesocyclone through its interaction with baroclinically generated horizontal vorticity produced by the cold pool (Wicker 1996). Alternatively, the prevalence of streamwise vorticity within the low-level wind profile could favor the development of a stronger low-level mesocyclone, which consequently enhances the upward-directed VPPGF responsible for converging and intensifying near-surface vertical vorticity (Markowski and Richardson 2009, 2014).

G18 investigated a less explored consequence of low-level shear, analyzing how shear affects supercell outflow position and effects on near-ground rotation. G18 performed a set of idealized supercell simulations with varying low-level shear properties, including depth, magnitude, and direction. They found that low-level shear orientation modulated the positioning of maximum near-surface circulation embedded within storm outflow relative to the midlevel mesocyclone, because of changes to low-level winds opposing the gust front. In particular, an easterly low-level shear profile resulted...
in the smallest displacement between these circulations, as well as the largest values of near-surface vertical vorticity. Though the favorability of this shear profile is not reflected in climatologies of tornadic supercell environments (e.g., Miller 2006; Esterheld and Giuliano 2008; Nowotarski and Jensen 2013), the notion that there exists an optimal positioning of a storm’s cold pool or low mesocyclone and its mesocyclone aloft for the maximum convergence and stretching of near-surface rotation is consistent with previous studies (Dowell and Bluestein 2002; Marquis et al. 2012; Skinner et al. 2014).

The simulations in G18 were initialized with the same thermodynamic profile, meaning that differences in outflow propagation were largely due to alterations of the low-level wind profile. However, given the expected relationship between LCL and outflow characteristics, it is plausible that LCL could influence mesocyclone positioning in a manner similar to that of low-level shear. We will address these issues by varying LCL in idealized simulations over the set of low-level wind profiles used by G18. Thus, we will explore how LCL influences cold pool properties, especially the positioning of circulation-rich outflow relative to the mesocyclone aloft, and production of near-surface vertical vorticity. Moreover, this study will contextualize the findings of G18 by examining how the thermodynamics and kinematics of the near-storm environment interact to affect outflow positioning and subsequent near-surface vertical vorticity intensification. This will ultimately provide insight into the physical mechanisms causing the observed relationship between LCL and tornadoes. Specifically, the following hypotheses will be tested by our experiments:

1) Changes in LCL will affect cold pool buoyancy in supercell thunderstorms, with higher (lower) LCLs leading to more (less) negatively buoyant outflow.
2) A lower LCL will lead to less forward propagation of outflow and its embedded near-surface circulation relative to the mesocyclone aloft in supercells compared to higher LCLs.
3) Near-surface vertical vorticity will be largest when the horizontal distance between the near-surface circulation and the mesocyclone aloft is minimized and the dynamic vertical perturbation pressure gradient force coincident with near-surface circulation is maximized.

While it is largely accepted that lower boundary layer RH leads to more outflow-dominated storms, the exact physical link between LCL and outflow buoyancy has yet to be tested rigorously due to the complicating influences of CAPE (e.g., McCaul and Cohen 2002; Lerach and Cotton 2012). Given that the last two hypotheses distinctly depend on outflow buoyancy, this link will be explicitly addressed through testing our first hypothesis. Section 2 describes the design of the modeling experiments performed, including the model configuration and the method employed to alter LCL in a way that minimizes CAPE differences. Section 3 presents the simulation results, addressing each hypothesis, and framing the results within the scope of existing literature and observations. Section 4 provides a summary and conclusions of our findings.

2. Methods and experiment design
   a. Numerical model configuration

   This study uses version 18 of Cloud Model 1 (CM1; Bryan and Fritsch 2002). CM1 is a three-dimensional, nonhydrostatic, fully compressible, time-dependent numerical model used to perform idealized simulations of a wide range of atmospheric phenomena (Bryan and Fritsch 2002). The same model configuration as G18 was implemented, in order for direct comparisons to be made between studies. The horizontal grid spacing is 500 m, with a stretched vertical grid of 50-m spacing below 4 km, increasing to 500-m spacing above 15 km. The model domain is 175 km × 175 km in the horizontal, and spans vertical depth of 20 km. Each simulation was run for 3 h. Adaptive time stepping is employed for the large model integration time step, with a smaller time step used to solve terms in the governing equations involving acoustic waves (Klemp and Wilhelmson 1978). Subgrid-scale turbulence is parameterized with 1.5-order closure (Deardorff 1980), and the Morrison double microphysics parameterization is used (Morrison et al. 2005). Rayleigh damping is applied above 15 km. Not included in these simulations are Coriolis force, radiation, and surface fluxes of heat, moisture, and momentum.

   b. Base states and initialization

   Convection is initiated using a warm bubble with a horizontal radius of 10 km and a vertical radius of 1.4 km, centered 1.4 km above the lower model surface at the horizontal center of the domain. After initialization, the model is integrated for 3 h. Approximately 1 h into integration for each simulation, a dominant right-moving supercell develops with a cold pool at the surface, which will be the focus of our analysis. Three thermodynamic soundings are used in this study, each with a different LCL. A base sounding corresponding to the thermodynamic profile used in G18 is computed using equations adapted from Weisman and Klemp (1982), with a surface pressure of 1000 hPa and an approximate LCL of 1 km. Though many methods
can be used to alter the LCL, such as changing the sub-LCL mixing ratio, this must be chosen carefully in order to limit the thermodynamic variability of the derived soundings. Namely, CAPE must not vary substantially as LCL is varied, as large differences could unduly affect convection and updraft intensity. A method was developed to alter LCL while minimizing CAPE differences, in which additional layers are added or removed from the base sounding and the sub-LCL mixing ratio is nudged such that a surface-based parcel follows a similar moist adiabat above the LCL in each sounding. This method is illustrated in Fig. 1a, and a sample model sounding is shown in Fig. 1b.

This process results in three thermodynamic soundings with different LCLs—approximately 0.5, 1, and 1.5 km—yet comparable CAPE values (the range of CAPE values is within 10% of their mean value). For the remainder of this study, these soundings will be...
referred to as our low-, medium-, and high-LCL cases (Fig. 1a). In addition to constraining CAPE differences, this method ensures that the thermodynamic structure above the LCL remains the same across all three soundings, eliminating variations in freezing levels relative to cloud base that could have undesirable impacts on hydrometeor distribution and buoyancy due to varied latent heating profiles. A necessary consequence of this method is that the surface pressure of the altered soundings vary from the base sounding (roughly +70 hPa for the higher LCL, and −80 hPa for the lower LCL). A number of sensitivity tests were performed to diagnose potential influences of these surface pressure variations on storm structure and outflow propagation. It was determined that the results of these tests were insensitive to changes in surface pressure. Precipitable water also varies between the soundings—with values of 1.13, 1.55, and 1.94 in. (1 in. = 2.54 cm) for the low, medium, and high LCLs, respectively. Table 1 contains additional information about the thermodynamic properties of the soundings.

A control run is performed for each LCL in which no additional shear depth is added at the bottom of the wind profile. The three remaining wind profiles are the same as those in G18, in which low-level shear depth, orientation, and magnitude are modified, while deep-layer wind profiles remain unchanged between simulations. This deep-layer profile consists of a clockwise-turning quarter-circle, with unidirectional westerly shear above 2 km (Weisman and Klemp 1984). For the low-level profile, a shear depth \( d \) of 500 m and shear magnitude of 7 m s\(^{-1}\) is chosen. Three low-level shear orientations \( \alpha \) (Fig. 1c), were tested—0° (easterly), 90° (southerly), and 180° (westerly). These four different wind profiles are applied across three different thermodynamic profiles with varying LCLs, for a total of 12 simulations. The names of the model simulations according to their LCL and low-level shear orientation are listed in Table 2.

The 0–3-km storm-relative helicity (SRH) values for the input hodographs are 247, 242, 357, and 252 m\(^2\) s\(^{-1}\) for the control, \( \alpha = 0° \), \( \alpha = 90° \), and \( \alpha = 180° \) hodographs, respectively. As noted in G18, the hodograph alterations were designed to minimize differences in SRH, in order to reduce the potential for changes in the strength of the low-level mesocyclone. However, as they discuss, some variability in SRH between wind profiles is unavoidable, as is particularly apparent in the \( \alpha = 90° \) simulations. Thus, our interpretation of results below must also consider changes in the strength of dynamic forcing provided by the low-level mesocyclones in each simulation.

### 3. Results

#### a. Thermodynamic characteristics of outflow as a function of LCL

Before any conclusions can be made regarding the effect of LCL on low-level rotation, we must determine how LCL influences the buoyancy of the outflow. As per our first hypothesis, we anticipate that there will be an inverse relationship between LCL and cold pool buoyancy, with higher LCLs leading to more negatively buoyant outflow. Two subdomains were defined for use throughout the proceeding analyses—a 50 km × 30 km

| Table 2. Simulation names by LCL and low-level shear angle \( \alpha \). |
|-----------------|-----------------|-----------------|
|                 | Low LCL (0.5 km) | Medium LCL (1 km) | High LCL (1.5 km) |
| \( \alpha = 0° \) | low_control      | med_control      | high_control     |
| \( \alpha = 90° \) | low_90          | med_90          | high_90         |
| \( \alpha = 180° \) | low_180        | med_180        | high_180       |

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1. To test storm sensitivity to varying surface pressure, a base sounding with Weisman–Klemp thermodynamic profile (Weisman and Klemp 1982) and a surface pressure of 1000 hPa is defined. This sounding is transformed to create two new soundings with identical temperature and relative humidity profiles as the base sounding, but with differing surface pressures of 950 and 1050 hPa. Model runs were initialized and run with these three soundings and the \( \alpha = 90° \) hodograph, and compared with one another. Additionally, 2D cold pool simulations were run with the same soundings and both the \( \alpha = 0° \) and \( \alpha = 90° \) hodographs to diagnose possible surface pressure influence on cold pool propagation. See supplemental Figs. 1–4 for results from these analyses.
domain box centered on the 2–5-km integrated updraft helicity maximum of the dominant right-moving supercell, and a smaller 25 km x 17.5 km box roughly centered on the updraft helicity maximum, which are referred to as the storm and inner domains, respectively. The former was chosen to capture the full structure of the analyzed storm, while the latter is meant to focus analyses on the portion of the cold pool and its embedded near-surface circulation that is most relevant to tornadogenesis. To determine the appropriate analysis times, time series of cold pool buoyancy within the inner domain were examined (Fig. 2). Buoyancy is an output variable from CM1 calculated using perturbation density potential temperature, with perturbations relative to the model base state. Negative values of buoyancy within our domain are assumed to be part of the cold pool. Cold pool buoyancy is similar across the LCLs until about 1.5 h into each simulation, at which point the buoyancy time series diverge to distinct values, particularly in the total and $\alpha = 90^\circ$ plots (Figs. 2a and 2c, respectively). This clear separation of the buoyancy profiles is an early indication that the high- (low-) LCL cases have the most (least) negatively buoyant outflow regardless of wind profile. Additionally, this divergence coincides with an increase in the fraction of the domain occupied by the cold pool (not shown), indicating when mature cold pools have developed in all of the simulations. Thus, all future analyses only consider the final 1.5 h of each simulation.

Plan views of buoyancy in the storm domain were generated for the $\alpha = 0^\circ$ and $\alpha = 90^\circ$ simulations (Figs. 3 and 4, respectively). Also included were contours of accumulated surface precipitation translated with the moving domain (shown in black), as well as the −1-K potential temperature contour (shown in dotted blue) which is used to estimate the position of the gust front. Though the control and $\alpha = 180^\circ$ simulations will be included in subsequent discussion, their storm structure will not be shown, since the control hodograph lacks low-level shear (which is atypical of supercell environments) and the $\alpha = 180^\circ$ simulations fail to produce appreciable near-surface vertical vorticity.

Both shear orientations exhibit similar temporal patterns across LCLs. In general, accumulated precipitation within the storm domain increases with time and is associated with an increasingly large and negatively buoyant cold pool, as is expected. The runs with a low LCL (Figs. 3a,d,g,j) display less negatively buoyant outflow relative the medium or high LCLs, though the low-LCL accumulated precipitation fields are not distinctly different than the other LCL configurations. The differences between the buoyancy and accumulated
precipitation fields in the medium- and high-LCL runs are more subtle, with these runs showing comparable magnitudes of buoyancy and accumulated precipitation, and the high-LCL case only containing a slightly broader area of negatively buoyant air by $t = 10800$ s (Fig. 3l). By $t = 9000$ s, there is a more negatively buoyant and broader cold pool in the high-LCL runs (Fig. 4i) relative to the medium-LCL runs (Fig. 4h). A general trend toward a more north–south-oriented gust front as LCL is increased is also evident (cf. Figs. 3j and 3l or Figs. 4g and 4i). Average gust front orientation was computed for each simulation by the methodology of G18 (section 2b), and the results (not shown) confirm this trend.

One explanation for these buoyancy trends could be differences in precipitation at and above the LCL, especially given that the precipitable water associated with each thermodynamic sounding actually increases with

![Fig. 3. Plan views of near-surface buoyancy (m s$^{-2}$) for the $\alpha = 0^\circ$ case in the storm domain, beginning (a)–(c) 1.5 h into model integration and shown (d)–(l) at half-hourly increments. Plotted data include accumulated precipitation translated with the moving domain (black contours plotted every 0.25 cm), the 2–5-km integrated updraft helicity maximum (black dot), and −1-K potential temperature perturbation (blue dotted contour; used to approximate gust front location). The units of the x and y axes are kilometers (from the UH max).]
increasing LCL. Distributions of rainwater mixing ratio $q_r$ at the LCL of each simulation during the final 1.5 h of model integration were analyzed (not shown). The control and $\alpha = 90^\circ$ simulations trend toward higher $q_r$ values with increasing LCL, but there is substantial overlap among the distributions. The $\alpha = 0^\circ$ and $180^\circ$ simulations lack such a trend, however, suggesting that differences in precipitation may be as sensitive to low-level shear properties as they are to precipitable water.

Therefore, absent a relationship between $q_r$ and LCL, the differences in cold pool buoyancy may be related to increased evaporation within the subcloud layer. Indeed, because the sub-LCL RH decreases as LCL is increased, the rain precipitating below cloud base falls into a progressively deeper and drier subcloud layer at higher LCLs, enhancing evaporation. This is evidenced by lower precipitation at the surface in the high-LCL case. This enhanced evaporation leads to the development of more negatively buoyant and faster-propagating cold pools, which is consistent with observations made from Figs. 2–4.

b. Effect of LCL on mesocyclone alignment

We can now explore how variations in cold pool buoyancy affect the positioning of near-surface circulation
relative to the mesocyclone aloft. Near-surface circulation is computed about a horizontal, 2-km-wide circuit centered on each grid point in the storm domain. By Stokes’ theorem, the circulation around a horizontal circuit is equal to the integral of vertical vorticity over the area enclosed by the circuit. Circulation thus provides a cumulative sense of vertical vorticity, as opposed to the more local view provided by vertical vorticity alone. Markowski and Richardson (2014) note that within this context, circulation acts as a measure of broad near-surface rotation that is available to be stretched by the mesocyclone aloft.

Figure 5 shows the circulation fields for the \( \alpha = 0^\circ \) simulations, at the same time steps as Figs. 3 and 4.\(^2\) For the low_\(\alpha_0\) simulation (left column), the near-surface maximum is located beneath the midlevel mesocyclone (approximated by the 500 m\(^2\) s\(^{-2}\) updraft helicity contour) at \( t = 5400 \) s (Fig. 5a). This circulation maximum shifts back and forth between this position and an area west of the mesocyclone aloft as areas of appreciable, positive circulation alternately develop and weaken in these locations. For the med_\(\alpha_0\) simulation (middle column), the near-surface circulation maximum is initially constrained behind the midlevel mesocyclone, becomes collocated with the updraft helicity maximum beneath the midlevel mesocyclone by \( t = 7200 \) s (Fig. 5e), and is located at the leading edge of the mesocyclone at both \( t = 9000 \) and 10,800 s (Figs. 5h and 5k, respectively). The high_\(\alpha_0\) simulation (right column) follows a similar pattern as the med_\(\alpha_0\) simulation, with the near-surface circulation maximum shifting eastward and ahead of the midlevel mesocyclone as time progresses. The only difference in the high_\(\alpha_0\) simulation is that the circulation maximum shifts further eastward (relative to the updraft helicity maximum) by \( t = 10,800 \) s (Fig. 5l), though the circulation remains in an area of appreciable helicity and strong midlevel updraft at this time.

To determine whether this relationship holds under a different low-level shear orientation, Fig. 6 shows the circulation fields for the \( \alpha = 90^\circ \) simulations.\(^3\) We can see similar patterns in the positioning of the near-surface circulation maximum as Fig. 5, but the eastward progression of this maximum is slowed. For example, the circulation maximum in the med_\(\alpha_90\) run remains west of the mesocyclone aloft by \( t = 7200 \) s (Fig. 6e) and becomes collocated with the midlevel mesocyclone starting at \( t = 9000 \) s (Fig. 6h), whereas the med_\(\alpha_0\) simulation has this collocation occurring at \( t = 7200 \) s (Fig. 5e) and shows the circulation maximum out ahead of the mesocyclone by \( t = 9000 \) s (Fig. 5h).

Though the preceding figures provide a general understanding of the positioning of near-surface circulation relative to the mesocyclone aloft, only so much information can be gleaned from these instantaneous fields. To assess the overall storm evolution, time series of maximum near-surface vertical vorticity and maximum integrated updraft helicity (used as a proxy for midlevel mesocyclone strength) are shown in Fig. 7, as well as two metrics meant to quantify the separation between near-surface circulation and the midlevel mesocyclone. The first is mesocyclone separation distance (MSD), which is defined in G18 as the distance between the updraft helicity maximum (i.e., the black dot in Figs. 5 and 6) and the near-surface circulation maximum (i.e., the green diamond in Figs. 5 and 6) within the inner domain. This quantity is positive (negative) when near-surface circulation is located west (east) of the mesocyclone aloft. Mesocyclone separation distance can vary rapidly (e.g., near the end of the high_\(\alpha_90\) time series; Fig. 7i) due to its sensitivity to areas of comparable near-surface vertical vorticity within close proximity of the updraft helicity maximum. Additionally, there is a tendency toward potentially unphysical, negative distances late in the simulations when the mesocyclone aloft becomes elongated or bifurcated (e.g., Fig. 6l). Therefore, time series of this quantity cannot be interpreted as following a single coherent mesocyclone center, and must instead be interpreted instantaneously.

Given the limitations of mesocyclone separation distance, a second metric referred to as average mesocyclone separation (AMS) was developed, which considers the separation between the centroids of appreciable near-surface circulation (circulation exceeding 5000 m\(^2\) s\(^{-1}\)) and the midlevel mesocyclone within the inner domain. The sign of this new quantity is determined by the positioning of these centroids in a reference frame aligned with the average gust front angle over the last 1.5 h of each simulation, rather than by their east–west displacement. These alterations result in more realistic values of mesocyclone separation, which are consistent with the computed circulation fields. Furthermore, AMS is only defined when appreciable circulation exists within the inner domain, thus removing separation values at times when there is no appreciable near-ground rotation to be intensified. Both separation metrics are shown in Fig. 7 for comparison.

First considering MSD, we see that for a given low-level shear orientation, the general trend of mesocyclone separation distances becomes more negatively sloped as LCL increases (e.g., in Figs. 7d–f). This implies that as LCL is increased and outflow becomes more

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\(^2\) See supplemental Fig. 5 for corresponding reflectivity fields.

\(^3\) See supplemental Fig. 6 for corresponding reflectivity fields.
negatively buoyant, near-surface circulation is generally transported farther forward relative to the midlevel mesocyclone. Additionally, the high LCL cases contain the lowest mesocyclone separation distances of any of the LCL configurations, for their respective low-level shear orientations. If we consider AMS, however, this trend is less clear with large values of MSD replaced with more moderate values of AMS.

The difference between MSD and AMS becomes even more evident in distributions of these metrics as shown in Fig. 8. Figures 8a and 8b show a general trend toward decreased mesocyclone separation distances as LCL is increased, consistent with Fig. 7, while Figs. 8c and 8d show minimal change in AMS as a function of LCL. One similarity between MSD and AMS, though, is that both show smaller distances for the $\alpha = 0^\circ$
simulations as compared with $\alpha = 90^\circ$ simulations. Also included in this figure are two new variables meant to further characterize the positioning of near-surface circulation. The first of these is the average circulation beneath the midlevel mesocyclone, which is a measure of the circulation-rich air near the surface available to be intensified by dynamic forcing provided by the mesocyclone. The second is a quantity referred to as circulation fraction, defined as the fraction of grid points beneath the low-level mesocyclone with near-surface circulation exceeding 5000 m$^2$ s$^{-1}$. This metric provides a slightly different perspective than the mesocyclone separation distance by quantifying the overlap of the low-level mesocyclone and areas of appreciable near-surface circulation, which is perhaps more relevant for the dynamic uplift and stretching of near-surface rotation. Regarding circulation, the $\alpha = 0^\circ$ simulations (Fig. 8e) display uniformly large and positive circulation beneath the midlevel mesocyclone for most of the final 1.5 h of the model run, while the $\alpha = 90^\circ$ simulations (Fig. 8f) generally contain small (and often net negative) near-surface circulation. In fact, the high$_{\alpha90}$ simulation is the only LCL configuration in Fig. 8f that contains circulation values on the same order.
of magnitude as any of the $\alpha = 0^\circ$ simulations. Circulation fraction decreases slightly with increasing LCL for the $\alpha = 0^\circ$ simulations (Fig. 8g), but as with average circulation, all three simulations realize larger overlap between the low-level mesocyclone and appreciable near-surface circulation than almost all of the $\alpha = 90^\circ$ simulations (Fig. 8h). The low_90 and med_90 simulations both contain minimal circulation fraction, with high_90 simulations being the only LCL configuration with larger values.

Figure 9 shows time series of average near-surface circulation beneath the midlevel mesocyclone and circulation fraction. For both the control and $\alpha = 90^\circ$ simulations, we see that the low-LCL configuration develops minimal, if any, circulation beneath the mesocyclone and displays nearly zero overlap between the low-level mesocyclone and appreciable near-surface circulation. As LCL is increased, both of these variables also increase, with the high_control and high_90 containing the highest average circulation and circulation fraction of their respective low-level shear orientations. That said, the increases in these variables within the aforementioned simulations do not occur until near the end of their model runs. The $\alpha = 0^\circ$ simulations show an entirely different evolution of these circulation quantities, with appreciable average circulation and circulation fraction in all three of its LCL configurations (with the largest values found in the
low_\alpha_0 case). In terms of timing, near-surface circulation and circulation fraction increase earlier than the other shear orientations, and remain high throughout the entire analyzed period. Additionally, the low_\alpha_0 and med_\alpha_0 cases contain larger values of both variables earlier than the high_\alpha_0, which shows more gradual increases.

All of the presented metrics suggest a complex relationship between LCL, low-level shear, and mesocyclone positioning. To visually confirm these quantitative results, heat maps of appreciable near-surface circulation were created. To do this, the computed circulation field at each time step during the final 1.5 h of model integration is remapped relative to the centroid of the midlevel mesocyclone. Points in this centroid-relative field with appreciable circulation are then counted across these time steps, and the resulting total frequency of occurrence at each location.

![Box-and-whisker plots](image-url)
is plotted in Fig. 10 with the average mesocyclone outline overlaid for reference. The control and $\alpha = 90^\circ$ simulations (Figs. 10a–c and 10g–i, respectively) both clearly show a more forward positioning of appreciable near-surface circulation relative to the mesocyclone aloft as LCL is increased. The $\alpha = 0^\circ$ (Figs. 10d–f) runs display substantial overlap between these features across all three LCLs, but the area of highest circulation counts still progresses out ahead of the midlevel mesocyclone with higher LCLs. Only the $\alpha = 180^\circ$ (Figs. 10j–l) runs lack such a trend, with the small amount of appreciable circulation realized in these simulations remaining well behind the mesocyclone aloft.

Both the MSD and AMS metrics paired with the qualitative interpretation of the circulation heat maps support our hypothesis that LCL influences the displacement of near-surface circulation relative to the mesocyclone aloft as LCL is increased. The $\alpha = 0^\circ$ (Figs. 10d–f) runs display substantial overlap between these features across all three LCLs, but the highest area of circulation counts still progresses out ahead of the midlevel mesocyclone with higher LCLs. Only the $\alpha = 180^\circ$ (Figs. 10j–l) runs lack such a trend, with the small amount of appreciable circulation realized in these simulations remaining well behind the mesocyclone aloft.

Revisiting Fig. 7, evidence of such a relationship can immediately be seen. There exists a general inverse correlation between our separation metrics and vertical vorticity across the simulations, with reductions in the magnitude of MSD and AMS roughly corresponding to increases in vertical vorticity, such as in low $\alpha$ (Fig. 7d) around $t = 8000$ s (for MSD) and $t = 8400$ s (for AMS). Conversely, for the $\alpha = 180^\circ$ simulations (Figs. 7j–l) in which both MSD and AMS remain large and positive throughout the analysis period, there is minimal production of near-surface vertical vorticity.

The connection between updraft helicity and near-surface vertical vorticity is less clear. Integrated updraft helicity remains relatively constant in all three of the control runs, suggesting that the strength of the midlevel mesocyclone does not vary rapidly with time. In both the $\alpha = 0^\circ$ and $\alpha = 90^\circ$ runs, there are increased updraft helicity values around times of increased near-surface vertical vorticity. That said, these increases in helicity tend to lag peaks of vertical vorticity, suggesting that they are a result of circulation-rich air being ingested into the storm, rather than the driving force behind the intensification of near-surface vertical vorticity. Last, this relationship could be influenced by differences in the SRH associated with each base state hodograph, and the fact that an integrated quantity such as updraft...
FIG. 10. Heat map of appreciable near-surface circulation within the inner domain relative to the midlevel mesocyclone centroid. The centroid-relative mesocyclone outline (i.e., the 500 m$^2$s$^{-2}$ contour) averaged over the last 1.5 h of model integration is outlined (green). The $x$ and $y$ axes are in kilometers (from the midlevel mesocyclone centroid).
helicity will sample different sections of the updraft (relative to its associated LCL) due to the LCL alteration method applied. The lack of a relationship between updraft helicity and near-surface vertical vorticity is significant, however, in that it suggests that the intensification of near-ground rotation is more a function of mesocyclone alignment than mesocyclone strength.

The importance of the low-level wind profile also is clear from Fig. 7. For $\alpha = 0^\circ$, the low-LCL case (Fig. 7d) corresponds to the largest peak vorticity values and the high-LCL case (Fig. 7f) contains the smallest vorticity values. For $\alpha = 90^\circ$, however, this relationship is reversed, with the high-LCL case (Fig. 7i) corresponding to the largest vertical vorticity values. Furthermore, the $\alpha = 0^\circ$ runs (Figs. 7d–f) contain the three largest values of all the simulations—which corroborates the surprising conclusions of G18 regarding the favorability of the $\alpha = 0^\circ$ wind profile—while the $\alpha = 180^\circ$ runs (Figs. 7j–l) contain three of the four lowest vorticity values.
This implies that certain low-level shear orientations may modulate the effects of varying LCL on storm dynamics, possibly by constraining or allowing the propagation of negatively buoyant outflow and low-level circulation beneath the midlevel mesocyclone.

To provide more definitive statistical support for our third hypothesis, the correlation between absolute MSD or AMS and near-surface vertical vorticity at each time step during the final 1.5 h of each simulation was computed. As noted by G18, the production of surface vertical vorticity does not instantaneously adjust to mesocyclone alignment, meaning that vertical vorticity values must be considered at some lag relative to these separation distances. A number of lag times were tested, including 0 (for means of comparison), 5, 10, and 15 min. Both 5- and 10-min lags yielded similar results, but a 5-min lag was chosen due to slightly higher correlation values. Figures 12a and 12b show a negative, statistically significant correlation between both MSD and AMS and vertical vorticity for each LCL consistent with the findings discussed thus far. Similar correlations exist when the results are separated by wind profile (not shown).

The influence of mesocyclone alignment can also be examined at lower levels in terms of circulation fraction, as defined in Fig. 8. Revisiting Figs. 7–9, we see that the simulations with more circulation located beneath the midlevel mesocyclone are associated with larger values of near-surface vertical vorticity. For example, the high_\alpha90 is the only \alpha = 90^\circ simulation containing large, positive near-surface circulation beneath its midlevel mesocyclone and appreciable circulation fraction (Figs. 8f,h and Fig. 9). Consequently, this simulation generates near-surface vertical vorticity values nearly twice that of the other LCL configurations (Figs. 7g–i). To this end, all three of the \alpha = 0^\circ simulations contain large average circulation and circulation fraction values, and subsequently develop intense near-surface vertical vorticity. These trends are also consistent with Fig. 12, as increases in both average circulation and circulation fraction generally correspond with periods of decreased mesocyclone separation (e.g., around t = 8000 s in Figs. 7d, 9a, and 9b).

Given these observations, the same lag correlations were performed with circulation fraction and near-surface vertical vorticity. As anticipated, Fig. 12c shows that there is a positive, statistically significant correlation between these variables, even when the points with a circulation fraction of zero are not included in the correlations. These points were excluded to avoid biasing our correlation values, though they reiterate the point that a lack of overlap between the low-level mesocyclone and near-surface circulation significantly limits the intensification of near-surface vertical vorticity.

Granted, the interpretation of these correlations is limited by the accuracy of MSD and AMS in characterizing mesocyclone alignment. Additionally, one must consider that near-surface vertical vorticity can often remain large even after MSD or AMS transition away from near-zero values, thus reducing the strength of these correlations. Though this increase in mesocyclone separation may seem unfavorable for strengthening of near-surface rotation, strong near-surface vertical vorticity may persist well after it is horizontally displaced from the midlevel updraft, much like in the findings of Dowell and Bluestein (2002). Thus, only a brief period of mesocyclone alignment seems necessary for initial near-surface vertical vorticity intensification, but not essential to its maintenance. In spite of these limitations, the presented time series (Figs. 7, 9), circulation heat maps (Fig. 10), and statistical analyses (Fig. 12) paint a consistent physical picture supporting the first half of our third hypothesis that near-surface vertical vorticity is maximized when there is overlap between near-surface circulation and the mesocyclone aloft.

Next, we will address the second half of the third hypothesis, investigating which physical mechanism is the driving force behind the observed relationship between mesocyclone separation and the production of near-surface vertical vorticity. For the purpose of this study, it is hypothesized that an upward-directed dynamic VPPGF attendant to the supercell mesocyclone is largely responsible for the stretching and intensification of near-surface vertical vorticity observed when mesocyclone separation approaches zero. As per the methodology of Hastings and Richardson (2016), which uses a similar approach as Rotunno and Klemp (1982, 1985), a hydrostatically balanced base state can be defined with \overline{\theta}, and \overline{\pi}—where \theta is density potential temperature and \pi = (p/p_0)\gamma cp is the Exner function—in which vertical accelerations are given by

$$\frac{Dw}{Dt} = -c_p \frac{\overline{\theta}}{\rho} \frac{\partial \pi'}{\partial z} + B,$$  \hspace{1cm} (1)

where w is the vertical component of velocity, c_p is the specific heat capacity at constant pressure, \pi' is the perturbation Exner function (\pi' = \pi - \overline{\pi}) which is akin to perturbation pressure, and B is the acceleration due to buoyancy. The terms on the rhs of Eq. (1) represent the contributions to vertical accelerations by VPPGFs and buoyancy, respectively. Within the cold pool of a supercell, the buoyancy typically provides a net negative
Therefore, it is assumed that dynamic VPPGFs are the primary drivers of low-level upward motion that can intensify near-ground vertical vorticity. However, it bears mentioning that in contrast to G18, buoyancy [in Eq. (1)] changes with LCL, so this must be taken into consideration when discussing the ability of dynamic vertical perturbation pressure gradients to lift circulation-rich outflow air.

The methods outlined in Hastings and Richardson (2016, their appendix A) were used to decompose the total nondimensional perturbation pressure $p'$ into its constituent parts—the buoyant $p_0^B$, linear dynamic $p_0^{LD}$, and nonlinear dynamic $p_0^{ND}$ perturbation pressures. As G18 notes, the $p_0^B$ term tends to be negative (positive) above (below) the local buoyancy minimum, $p_0^{LD}$ is positive (negative) upshear (downshear) of updrafts and negative (positive) upshear (downshear) of downdrafts, and $p_0^{ND}$ is positive in regions of deformation and negative in regions of vorticity. Though $p_0^B$ is typically positive within negatively buoyant outflow near the surface, its associated pressure gradient force is generally counteracted by the buoyancy term in Eq. (1). Thus, the dynamic part—meaning the sum of the linear and nonlinear dynamic vertical perturbation pressure gradient forces—will be the focus of our discussion. Specifically, the dynamic VPPGF five minutes before maximum near-surface vertical vorticity will be analyzed, given the hypothesized delayed response between dynamic forcing and responses in the vorticity field (as per Fig. 12).

Figure 13 shows the low-level (500 m AGL) dynamic VPPGF field for each LCL in the $\alpha = 0^\circ$ and $\alpha = 90^\circ$ cases at the time of maximum near-surface vertical vorticity, with contours of the midlevel mesocyclone and near-surface circulation superimposed. As expected, the
midlevel mesocyclone is associated with positive low-level (500 m AGL) dynamic VPPGF leading up to the time of maximized near-surface rotation. Furthermore, some overlap exists between the dynamic VPPGF and positive near-surface circulation prior to the maximization of near-surface vertical vorticity.

To identify any trends in the decomposed VPPGFs, Fig. 14 shows the average near-surface circulation and components of Eq. (1) beneath the midlevel mesocyclone computed at the times shown in Fig. 13. All vertical accelerations are averaged over the lowest 500 m of the model domain in order to approximate the low-level uplift responsible for the stretching and intensification of near-surface vertical vorticity. One observation from Fig. 14 is that for both the $\alpha = 0^\circ$ (Fig. 14b) and $\alpha = 90^\circ$ (Fig. 14d) simulations, the accelerations due to buoyancy and buoyant perturbation PGF are of roughly equal and opposite magnitude, leading to a minimal total buoyant forcing. Moreover, the nonlinear VPPGF dwarfs the linear VPPGF (as noted in G18), and therefore controls the magnitude of the dynamic VPPGF.

In the $\alpha = 0^\circ$ simulations, the low LCL configuration (which produces the largest near-surface vertical vorticity; cf. Fig. 7) contains the largest near-surface circulation and dynamic VPPGF collocated beneath the midlevel mesocyclone (Figs. 14a,b). As LCL increases, the dynamic VPPGF decreases in magnitude such that it is almost entirely canceled out by the increasingly negative total buoyant forcing. This supports the hypothesis that environmental RH (and by extension, LCL) affects the uplift of outflow parcels by altering their buoyancy (e.g., Markowski and Richardson 2009, 2014). For the $\alpha = 90^\circ$ simulations, the medium LCL configuration contains the largest dynamic VPPGF. However, this forcing is located above an area of negative circulation (Fig. 14c), limiting the intensification of near-surface vertical vorticity. The simulation with the second highest dynamic VPPGF—the high LCL configuration—is coincident with appreciable, positive near-surface circulation, and therefore can contribute to the development of the most intense near-surface rotation of the $\alpha = 90^\circ$ simulations. This occurs in spite of quite negative buoyant forcing, which lends supports to the argument that positioning of forcing relative to near-surface circulation is also crucial for the stretching and intensification of near-surface vertical vorticity. Thus, it appears that both outflow buoyancy and mesocyclone positioning contribute to the magnitude of rotation realized at the surface within supercells.

d. Additional considerations

Though the focus of this study is the influence of LCL on near-surface outflow and rotation, it is worth noting the sensitivity of the presented results to low-level shear orientation. As discussed previously, the LCL corresponding to the maximum near-surface vertical vorticity varies depending on the low-level shear orientation. The average gust-front orientations were also used to compute the magnitude of storm-relative winds opposing the gust front in each simulation. These calculations combined with the observations from Fig. 7 reveal that the $\alpha = 0^\circ$ profile contained the smallest component of low-level, storm-relative winds opposing the cold pool, and generated the largest values of near-surface vertical vorticity when paired with a low LCL. Conversely, the shear profiles with moderate winds opposing the cold pool—control, $\alpha = 90^\circ$—maximize their vertical vorticity with a high LCL. Last, when there are large gust-front-opposing winds ($\alpha = 180^\circ$), the production of near-surface vertical vorticity is limited across all LCLs.

These insights reveal two surprising results. First is the favorability of the $\alpha = 0^\circ$ wind profile across all three LCLs. Previous studies (e.g., Wicker 1996; Miller 2006; Esterheld and Giuliano 2008; Nowotarski and Jensen 2013; Coffer and Parker 2017) have identified this shear profile as being unfavorable for tornadoogenesis. Specifically, this hodograph shape is associated with substantially crosswise vorticity at lower levels, weaker gust-front-opposing winds, and easterly shear which favors outflow-dominated vorticity in which appreciable near-surface circulation surges out ahead of the mesocyclone aloft (e.g., Frame and Markowski 2013; Nowotarski and Jensen 2013; Parker 2014). However, the $\alpha = 0^\circ$ simulations correspond to the three largest near-surface vorticity maxima of all the simulations performed. This result, though counterintuitive, is in agreement with the findings of G18 in which the $\alpha = 0^\circ$ wind profile was the only low-level shear orientation which led to the formation of intense near-surface vertical vorticity. Similar to the reasoning presented by G18, decreased gust-front-opposing winds in this case actually facilitate a rapid decrease of MSD and near-zero AMS (cf. Figs. 7d–f), allowing for the alignment of the maximum dynamic low-level updraft forcing and near-surface circulation. This is evidenced by Figs. 10d–f, with a higher frequency of appreciable circulation beneath the midlevel mesocyclone in the $\alpha = 0^\circ$ simulations than the other shear orientations, regardless of LCL. The reason for these unexpected results is admittedly unclear, and could be related to factors not addressed by this study such as friction or storm sensitivity to the deep-layer shear profile. Within the context of this study, however, the important finding is that the simulations that minimize...
mesocyclone separation realize the largest values of near-surface vertical vorticity.

The second intriguing result is that low_α90 did not contain the largest vertical vorticity values of the α = 90° simulations. Given the storm climatologies discussed, it would be expected that this environment would be prime for the formation of intense near-surface vertical vorticity, given its low LCL and ample streamwise vorticity associated with its hodograph shape. However, low_α90 (cf. Fig. 7g) corresponds to the lowest near-surface vertical vorticity of the α = 90° simulations, with high_α90 (cf. Fig. 7i) producing near-surface vertical vorticity over twice as strong. This is corroborated by the high_α90 circulation heat map (cf. Fig. 10i) which shows higher counts of appreciable circulation beneath the midlevel mesocyclone, in comparison with the low_α90 heat map. The physical explanation for this occurrence may be the same as that
offered earlier with the \( \alpha = 0^\circ \) results. Additionally, it is possible that the model configuration itself (grid spacing, microphysics parameterization, etc.) could influence the favorability of one LCL–hodograph combination over another. Again though, as with the \( \alpha = 0^\circ \) simulations, these results highlight the importance of mesocyclone alignment for the intensification of near-surface rotation.

At face value, these unexpected findings could bring into question the physical relevance of the results presented. It is therefore important to remind the reader that given the constraints of using a numerical model, the intent of this study is not to identify specific combinations of LCL and low-level shear orientation that favor tornadogenesis. Rather, our goal is to examine the sensitivity of the storm morphology and maximum near-surface vertical vorticity to these parameters, so as to highlight the basic physical mechanisms that might contribute to the development of near-surface vertical vorticity. In reality, the existence of heterogeneity in the near-storm environment, interactions between outflow from different storms, and complex microphysical processes, among other things, add a layer of complexity to the process of tornadogenesis that extend beyond the scope of this study.

4. Summary and conclusions

The impact of varying LCL on outflow characteristics and the production of near-surface vertical vorticity was tested through a set of idealized supercell simulations. Regarding our first hypothesis, there is a distinct tendency toward more negatively buoyant cold pools as LCL is increased (Figs. 2–4). Given the lack of a consistent trend in \( q_r \) values across the three LCLs, as well as the decreasing of subcloud RH with increasing LCL, it is likely that these observed variations in outflow buoyancy are related to increased evaporation of rainfall in environments with higher LCLs.

In terms of our second hypothesis, results (Figs. 7–11) indicate that higher LCLs correspond to a more forward positioning of appreciable near-surface circulation relative to the mesocyclone aloft. This is physically consistent with the results from our first hypothesis, in that more negatively buoyant outflow associated with higher LCLs allows for faster transport of near-surface...
circulation, which as a result is more likely to advect beneath and subsequently ahead of the mesocyclone aloft. Additionally, the magnitude of this mesocyclone separation appears to be sensitive to variations in the low-level wind profile, which in turn influences the magnitude of storm-relative winds opposing the gust front.

The first part of our third hypothesis regarding the importance of circulation overlap has been demonstrated both in this study, and by G18. The time series presented in Fig. 7, as well as the correlations and analysis in Fig. 12, demonstrate that decreased distance between the near-surface circulation and mesocyclone aloft, or similarly increased overlap between the low-level mesocyclone and near-surface circulation, leads to an intensification of near-surface vertical vorticity. The second portion of the hypothesis was validated by decomposing the simulated pressure fields and analyzing the strength and positioning of the vertical accelerations beneath the midlevel mesocyclone (Figs. 13, 14).

In the analyzed simulations, there is a positive dynamic VPPGF beneath the midlevel mesocyclone prior to the maximization of near-surface vertical vorticity. However, large dynamic VPPGF does not guarantee the production of intense vertical vorticity, as this forcing cannot fully contribute to the strengthening of near-surface vertical vorticity until it is collocated with appreciable, positive circulation at the surface. Additionally, more negative buoyant forcing with higher LCLs can negate the positive VPPGF, limiting the stretching of near-surface rotation.

Future work is necessary to fully understand the influence of LCL on supercell evolution and the formation of intense near-surface rotation. Perhaps most importantly, the findings of this study and G18 must be compared with the near-storm environments of observed tornadic supercell cases, particularly those with higher LCLs and/or an ambient wind profile similar to our $\alpha = 0^\circ$ hodograph. These conditions were found to favor the production of intense near-surface vertical vorticity in several of our simulations, despite storm climatologies indicating that low LCLs and an ambient shear profile with ample streamwise vorticity (e.g., the $\alpha = 90^\circ$ wind profile) are associated with tornadic supercells. Thus, it would be of interest to determine whether the storm environments that produced intense near-surface vertical vorticity are found in nature or if some other physical mechanism can explain the divergence of our simulation results from observed storms. Another unexplored consequence of altering LCL in the manner prescribed within this study is that it changes the height of LCL relative to the shear profile. One can imagine how this might affect the spatial distribution and size sorting of hydrometers, or the amount of ambient horizontal vorticity ingested and tilted by the updraft. Such alterations could thus influence the observed patterns in accumulated surface precipitation, and by extension the buoyancy and orientation of storm outflow or the strength of the low- and midlevel mesocyclone.

Overall, this study has provided insight into the impact of varying low-level humidity on outflow and rotation within idealized supercell thunderstorms while maintaining a roughly constant buoyancy profile. Though there is not one particular LCL that always favors the development of intense near-surface rotation, it is clear that the LCL has an influence on the outflow characteristics in simulated supercells. These outflow characteristics, in combination with the low-level wind profile, dictate the relative positioning of midlevel and near-surface circulation, which in turn regulates the production of near-surface vertical vorticity. Thus, the LCL can be said to play a nuanced, yet crucial role in determining whether intense surface rotation is able to form within simulated supercells. Depending on the low-level wind profile within a given storm environment, the most favorable LCL for tornadogenesis may not always be the lowest.

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