THE CHARACTERSISTICS OF NUMERICALLY SIMULATED
SUPERCELL STORMS SITUATED OVER STATICALLY STABLE
BOUNDARY LAYERS

A Thesis in
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by
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Abstract

Numerical simulations of supercell thunderstorms are used to assess changes in vertical motion and low-level rotation in environments with differing low-level static stability. Simulations are initialized in an idealized, horizontally homogeneous environment with a shallow stable layer that is representative of a nocturnal inversion or a mesoscale cold pool. The depth and temperature deficit of the imposed stable boundary layer, which together define the convective inhibition (CIN), are varied in a suite of simulations.

When compared with a control simulation with little surface-based CIN, each supercell simulated in a stable boundary layer exhibits weaker low-level vertical vorticity and weaker low-level vertical velocity despite similar most unstable convective available potential energy (MUCAPE); in general, low-level vertical vorticity and vertical velocity decrease as CIN increases. It was found that while the presence of a stable boundary layer decreases low-level updraft strength, all supercells except those initiated over the most stable boundary layers had some updraft parcels with surface origins. Furthermore, the existence of a stable boundary layer does not prohibit downdraft parcels from reaching the surface, though decreased negative buoyancy decreases downdraft speed. Trajectory and circulation analyses indicate that the weaker rotation at low levels is a result of the decreased generation of circulation at low levels coupled with decreased convergence of the near-ground circulation by weaker storm updrafts in the stable-layer scenarios. These results also may suggest a reason why tornadogenesis is less likely to occur in elevated supercell thunderstorms.
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Chapter 1

Introduction and Background

Severe convective storms have been the subject of a large body of atmospheric science research, no doubt in large part owing to the threats they pose to society. Tornadoes, in particular, have received much of the attention, although other hazards of convective storms such as flooding, hail, straight-line winds, or lightning are more common and are cumulatively more destructive to life and property.

Typically, the most intense subset of tornadoes are spawned within the class of thunderstorms referred to as supercells. Favorable environments for supercells are characterized by large vertical wind shear over a deep layer of the troposphere, coupled with ample convective available potential energy (CAPE). The likelihood of tornadogenesis is increased in supercell environments that also are characterized by strong low-level shear and a relatively humid boundary layer. In contrast, some environments support supercells yet have other attributes that seem to suppress tornadoes. For example, statically stable boundary layers are considered to inhibit tornadogenesis (Davies 2004). In such a situation, shear and CAPE may be in abundant supply, but the low-level dynamics responsible for tornadogenesis are likely hampered by relatively cool air present at the surface. Any effort to better predict tornadoes and the broader low-level circulations that precede their development must examine the causes of tornadogenesis failure where the environment might otherwise be favorable for supercells.

In this study, the effects of low-level static stability on the characteristics of supercell thunderstorms are examined using numerical simulations. One limitation of the simulations is that they only relate statically stable boundary layers to
the intensity of low-level rotation on a scale resolved by the model grid, which is too coarse to resolve tornadoes. The relationship between tornado formation and the intensity of low-level rotation on a larger scale (on the order of 1 km) is unfortunately a bit temous. It would seem likely that strong low-level rotation on such a larger scale would be a prerequisite for tornadogenesis. However, strong rotation on scales larger than the tornado does not ensure tornadogenesis. In fact, it is possible in some cases to have too much circulation at low levels (the radially outward-directed centrifugal force associated with a rotating flow increases with circulation and could prevent vortex contraction/intensification in a case with very strong circulation, such that the centrifugal force exceeds the radially inward-directed pressure gradient force; e.g., Markowski et al. 2003). Therefore, the presence of strong low-level rotation is required for tornadogenesis but does not always indicate that a tornado will form.

1.1 Airflow and Vorticity Generation within Supercell Thunderstorms

For nearly fifty years, the term “supercell” has been used to describe a particular class of organized and often severe convection. Browning (1964) first proposed a kinematic model (Fig. 1.1) of the tornadic supercell thunderstorm with later alterations by numerous others including the often cited model (Fig. 1.2) put forth by Lemon and Doswell (1979). Defining features of supercell thunderstorms include a rotating updraft (or mesocyclone) driven by both buoyancy and an upward-directed vertical dynamic perturbation pressure gradient force, as well as two downdraft regions with their attendant gust fronts.

A supercell updraft acquires midlevel rotation (i.e., its midlevel mesocyclone) as ambient horizontal vorticity associated with environmental vertical wind shear is tilted and stretched vertically by the updraft (Davies-Jones 1984; Lilly 1986). As vertical vorticity increases with height, a negative vertical perturbation pressure gradient results due to the dynamic lowering of pressure associated with rotation. This dynamically-induced vertical pressure gradient enhances the updraft, and is especially critical at low levels, where even stable air that originates in
the precipitation-cooled outflow of the storm is forcibly lifted to its level of free convection (LFC) (Marwitz 1973; Bonesteelle and Lin 1978). In many supercell environments, this vertical pressure gradient is enhanced by a varying orientation of the environmental shear vector with height. Because dynamic high (low) pressure develops upshear (downshear) of the maximum updraft, changes in the direction of the shear vector with height create a dynamic vertical perturbation pressure gradient. For a vertical wind profile in which the shear vector veers with height (such as that used in the present simulations) this effect favors a cyclonic right-moving storm as discussed in detail by Klemp and Wilhelmson (1978a) as well as Rotunno and Klemp (1982).

![Figure 1.1. Schematic diagram of airflow within a right-moving supercell thunderstorm. Several precipitation trajectories (dotted paths) as well as the surface precipitation field (shading) are shown. The surface gust front and typical position of a tornado are also shown. [From Browning 1964, his Fig. 2]](image)

The updraft and downdraft structures of supercell thunderstorms have been well-documented by previous studies. Diminished radar reflectivity within an area of the storm with strong upward motion, known as the weak echo region (WER), suggests that the main updraft is relatively precipitation free (Browning and Donaldson 1963; Browning 1964, 1965; Auer and Sand 1966). Goldman (1968) proposed a conceptual model of storm airflow (Fig. 1.3) in which the updraft air is
largely supplied by an inflow layer at the surface. As the updraft rises, dry environmental air is entrained at low and midlevels. Precipitation falls out of the sloped updraft into unsaturated adjacent air which is evaporatively cooled. This process creates negative buoyancy and leads to the formation of a downdraft as described by Browning and Ludlam (1962) and Brandes (1982). Downdrafts might also be initiated by midlevel environmental flow impinging upon the updraft. Barnes (1978) found that as the flow decelerates at a stagnation point in contact with the updraft, dynamic forcing in combination with evaporative cooling causes descent. It is likely that a combination of these processes, rather than just one acting alone, results in the supercell downdraft structure generalized in Fig. 1.2. The often contiguous downdraft has maxima on both the forward and rear flanks of the storm. The forward-flank downdraft is primarily due to the evaporation of hydrometeors falling downstream from the updraft. The rear-flank downdraft may be forced by both the thermodynamic and dynamic processes described above, but the relative

Figure 1.2. Schematic diagram of surface characteristics of a supercell thunderstorm. The rear- and forward-flank downdrafts (RFD, FFD) and updraft (UD) regions are shown. The position of the surface gust front, surface airflow and likely tornado location are also shown. [From Lemon and Doswell 1978, their Fig. 7]
importance of each process remains unclear (Markowski 2002).

**Figure 1.3.** Conceptual drawing of Goldman’s airflow model in thunderstorms (a) viewing northward and (b) southeastward for eastward motion. [From Goldman 1967, his Fig. 3]

**Figure 1.4.** (a) Barotropic, ambient vorticity is reoriented by an updraft. Whereas midlevel vertical vorticity results from this process, tilting cannot occur at the ground, thus vertical vorticity cannot occur there by means of an updraft alone. (b) In the presence of both an updraft and downdraft, vertical vorticity is produced near the ground through tilting by a downdraft and then stretched by the updraft at low levels. [From Markowski et al. 2008, their Fig. 16].

In cases where no prior low-level vertical vorticity exists, previous investigators have postulated that both the forward- and rear-flank downdrafts are necessary ingredients in generating circulation near the ground. Rotunno and Klemp (1985) found that the cold pool created by the forward-flank downdraft leads to the baroclinic generation of horizontal vorticity along the forward-flank gust front which then may be drawn into the low-level updraft. However, Davies-Jones and Brooks
(1993) argued that tilting and stretching of low-level horizontal vorticity by an updraft alone cannot produce a tornado because horizontal vorticity cannot be tilted at the ground by an updraft. Thus, without a downdraft there is no vertical vorticity at the ground that can be amplified through stretching (Fig. 1.4a). They found that positive near-ground vertical vorticity is a result of the positive tilting of vorticity within descending air. This process is illustrated as the downward bending of horizontal vortex lines in Fig. 1.4b. The resulting parcels with cyclonic vorticity are then advected into the updraft from its southwest side where further tilting and convergence amplifies their vorticity. Markowski et al. (2008) found that vortex lines passing through low-level maxima in vertical vorticity had arching structures which they concluded were likely to have resulted from the baroclinic generation of vorticity that was then tilted by an updraft, in contrast to a simple redistribution of environmental vorticity. Walko (1993) also verified the importance of a downdraft in tilting vorticity near the surface in numerical simulations. Additionally, observations that link tornadoes with the presence of hook echoes, clear slots, and their associated rear-flank downdrafts support the notion that the rear-flank downdraft is related to tornadogenesis (see Markowski 2002). Though the difference in the roles played by the forward- and rear-flank downdrafts in tornadogenesis remains ambiguous, evidence suggests that baroclinic vorticity generated by the horizontal buoyancy gradients resulting from one or both downdrafts often, if not always, is an important contributor to tornadogenesis in the absence of pre-existing vertical vorticity.

1.2 Elevated Supercells

There is considerable anecdotal evidence that supercell thunderstorms are sensitive to changes in the environmental low-level static stability and convective inhibition (CIN). For example, “elevated” supercells are widely assumed to pose a greatly diminished tornado threat compared to “surface-based” supercells. [An elevated storm (Colman 1989) is defined herein as one that draws its inflow from a layer not in contact with the surface because of the presence of an underlying, surface-based, cold air mass that is independent of the storm’s precipitation region.] Environments containing a boundary layer characterized by relatively large surface-based
CIN are conducive to elevated supercells because in such cases the dynamic vertical pressure gradient within a storm may not be able to lift negatively buoyant air parcels from the surface to their level of free convection (LFC). Thus, storm inflow could not originate from the surface directly below the storm. Elevated storms are most often observed above stable nocturnal boundary layers and on the cool side of a thermal boundary (e.g., a front, outflow boundary, or other mesoscale air mass boundary associated with a temperature gradient). In simulations using an axisymmetric model, Leslie and Smith (1978) demonstrated “the likely importance of static stability as a factor opposing tornadogenesis.”

A major nowcasting challenge exists in assessing the likely changes in rotation for a storm crossing a thermal boundary because the low-level vertical wind shear is typically enhanced on the immediate cool side of the boundary owing to the baroclinic generation of horizontal vorticity. Thus, there are potentially competing effects in that the increased CIN tends to be detrimental to low-level rotation, whereas the increased low-level shear tends to favor low-level rotation. For example, Maddox et al. (1980), Weaver and Purdom (1995), and Rasmussen et al. (2000) documented a significant number of tornadoes within supercells shortly after the storms crossed thermal boundaries. Furthermore, Markowski et al. (1998) found that almost 70% of tornadic storms observed during the 1995 VORTEX field campaign were associated with pre-existing mesoscale boundaries. Numerical simulations by Atkins et al. (1999) further support the notion that horizontal vorticity enhancements along such boundaries are important.

Alternatively, others have documented supercells that were tornadic or had rapid intensifications of low-level rotation while interacting with a thermal boundary, but then became elevated and nontornadic after crossing the boundary into air masses characterized by substantial surface-based CIN. Markowski et al. (1998) found that within the subset of VORTEX tornadoes associated with preexisting boundaries, most tornadoes in the cooler air mass occurred within 30 km of the relevant boundary. It is likely that storms that moved too far across the boundary into a cool air mass encountered too much surface stability to sustain tornadoes.

In a specific case study, Doswell et al. (2002) examined the 7–8 June 1998 event in eastern New Mexico and western Texas in which a cloud-covered region of cool air associated with the outflow of prior thunderstorms may have limited
the tornadic potential of a supercell. Whereas the storm was briefly tornadic while interacting with the boundary of the cool air, tornadic potential diminished upon crossing into the low-level stable region despite the storm persisting as a supercell. Reports of large hail continued even after the storm crossed into the cloud-covered region, yet strong surface wind gust reports ceased. This suggests that although the storm was able to maintain a vigorous updraft, even while situated over the cool air mass, downdrafts were weakened as they approached the surface by the statically stable boundary layer.

Similarly competing shear and buoyancy effects are commonly observed when nocturnal boundary layers form, as the low-level stabilization tends to be accompanied by the development of a nocturnal low-level wind maximum and an increase in low-level shear. Developing an understanding of the dynamics that suppress low-level rotation in the present simplified numerical experiments is a first step toward improving the ability to anticipate how a storm will respond to the static stability and vertical wind shear modifications associated with nocturnal cooling or a mesoscale-boundary crossing.

1.3 Hypotheses and Experiment Design

Owing to the scarcity of complete observational datasets of supercell thunderstorm wind and thermodynamic fields as well as the desire for control over experimental parameters, the present study utilizes an idealized numerical modeling approach similar to that which has been popular for a number of other investigations of supercell dynamics (e.g., Klemp and Wilhelmson 1978; Rotunno and Klemp 1982, 1985; Weisman and Klemp 1982, 1984). In this study, supercells are simulated within horizontally homogeneous environments having varying degrees of surface-based CIN. This research studies the (presumably) simpler case of a supercell occurring in a nocturnal boundary layer rather than the case of a supercell encountering horizontal heterogeneity in the form of a preexisting thermal boundary. The supercells simulated over a stable boundary layer are compared to a supercell in a control experiment initialized in an environment with negligible low-level static stability (the control case is initiated with weak static stability to ensure the environmental Richardson number is less than 0.25). The a priori assumption is
that the introduction of a strongly stratified layer in contact with the surface will force the simulated storms to draw inflow from a higher altitude than the control; however, as will be seen later, this presumption may not always be valid owing to the strong upward-directed dynamic vertical pressure gradient forces present in supercells. Thus, the first goal of this study is to elucidate variations in updraft and downdraft airflow within a range of stable layer characteristics. It may be that storms above weaker stable layers are still able to draw inflow air from the surface directly below, thus remaining surface-based. Furthermore, it is possible that a storm may not become elevated yet still exhibit weaker low-level rotation.

Beyond investigating the influence of the low-level stratification on the vertical excursions of parcels, this research is geared toward answering the following question: “Why are elevated supercells less likely to produce tornadoes than surface-based supercells?” The goal is to identify the dynamics responsible for the suppression of low-level rotation in the elevated supercells. As previously discussed, though the simulations are unable to resolve tornadoes, it is inferred that a decrease in low-level rotation would be detrimental to tornadogenesis. The simulations are designed to determine if the suppression of near-ground vertical vorticity in elevated storms owes to a lack of near-ground circulation as downdrafts are less able to penetrate the stable air mass (recall from Section 1.1 that downdrafts are the only means by which vertical vorticity can develop at the surface in a horizontally homogeneous environment without a Coriolis effect), inhibited convergence of near-ground circulation as a result of weaker vertical velocities, or a combination of both effects.

The configuration of the model, environmental parameters, and analysis methods are further discussed in Chapter 2. Chapter 3 summarizes the results of the parameter space investigation; that is, the relationship between the various storm environments and the characteristics of the updrafts, downdrafts, and mesocyclones at low-levels and midlevels. Discussion of results and a more detailed comparison of the control supercell with two simulations having moderate and strong low-level stratification aimed at investigating the dynamics responsible for the differences in near-ground rotation are presented in Chapter 4. Chapter 5 provides a summary and the conclusions drawn from the simulation results.
Chapter 2

Methods

Numerical modeling is implemented as a method of isolating the effects of variations in environmental low-level static stability on supercell thunderstorms. Though any numerical investigation is inherently constrained by the limitations of the model itself, simulation often provides a useful alternative to observations. The dataset generated by such a model is more comprehensive than any collection of observed data. With full knowledge of the state variables, moisture, and wind fields, the modeler is able to systematically analyze any aspect of the simulated storm’s structure and evolution. Furthermore, it stands to reason that any feature observed in the model output can be accurately explained using the model-derived variables and knowledge of the model processes. Another unique advantage of modeling is the flexibility in adjusting any one environmental parameter as a means of isolating its effect on the simulation results. Finally, numerical modeling enables the investigation of a full parameter space of simulations initiated with environmental characteristics that datasets of observed storms may lack.

Given the seemingly unlimited potential of modeling in answering research questions, it may be easy to quickly accept all simulation results as fact. However, one must take great care to interpret model results in the proper context. A storm produced in a model is only a solution predetermined by a set of discretized equations and initial conditions. The results of a numerical simulation may vary depending on any number of factors including parameterization of physical processes, advection schemes, boundary conditions, or the discretization method. Despite these caveats, a great deal can be learned about the physical world through
modeling. In a study such as this, where the sensitivity of a supercell to only one environmental parameter is in question, numerical modeling is an appropriate method of inquiry. In order to isolate the effects of low-level static stability on low-level rotation it is necessary to eliminate other environmental parameters that affect low-level rotation and may also vary with the addition of a stable boundary layer (e.g., low-level shear). It is likely that purposely neglecting realistic processes skews some aspects of the simulation results from what might be observed in nature. However, the effects of the variable of interest are presumed to reflect how modifications of only that variable might affect a real storm. Therefore, the negative correlation between low-level static stability and near-ground rotation investigated herein as well as its underlying causes can be confidently extended as an explanation of observation.

In addition to directly comparing model simulations as a means of assessing supercell sensitivity to varying environments, further analysis is made possible by the full dataset of the modeled wind field. This study relies heavily on parcel trajectories integrated from the wind fields to gain understanding of both the structure of vertical airflow in elevated supercell thunderstorms and the dynamics governing decreased low-level rotation in such storms. In the latter case, trajectories are a central part of the circulation analysis that will be further detailed in Chapter 4.

2.1 Numerical Model Configuration

The numerical simulations were performed using the Bryan Cloud Model (CM1), Version 1, Release 11 (Bryan and Fritsch 2002; Bryan 2002). CM1 is a moist, non-hydrostatic model well-suited for idealized research of mesoscale convective phenomena. The model solves the compressible governing equations, using a split time step to solve for those terms associated with acoustic waves using a smaller timestep than the remaining terms, following the technique of Klemp and Wilhelmson (1978b). In this configuration, a large integration time step of 3 s is used with a smaller time step of 0.5 s. The governing equations are integrated using a Runge-Kutta time differencing technique, and horizontal and vertical advection are computed using fifth-order spatial derivatives with implicit diffusion. As such, no artificial diffusion is necessary. Each simulation covers a 2-hour time period.
This integration time allows for sufficient “spin-up” time for each storm, as well as enough time in the mature phase to objectively compare simulations at similar phases in their life cycle. The model simulations neglect the effects of radiation as well as surface fluxes of momentum, moisture, and heat. The Coriolis force is also ignored because its effects are negligible over the relatively short (2 hour) simulation time. Although these physical processes would likely influence the simulation results, their inclusion would lead to a changing base state. The goal of these experiments is to isolate the effects of a constant stable boundary layer on storm evolution, therefore a base state that evolves throughout each simulation would unnecessarily complicate the analysis.

The model uses the ice phase microphysical parameterizations developed by Lin et al. (1983), allowing for the simulation of water vapor, cloud water, cloud ice, rain, snow and hail using a single moment (prognosing only mixing ratio) bulk water microphysical parameterization. Though hydrometeors are not of any direct concern to this investigation, this scheme was chosen over a more simple warm-rain scheme because thermodynamic properties of the modeled thunderstorm downdrafts and cold pools have shown sensitivity to the microphysical parameterization (Gilmore et al. 2004; Snook and Xue, 2008). Considering the presumed importance of the downdraft and cold pool characteristics in generating low-level vertical vorticity, it is a priority to model hydrometeors as realistically as possible in this study. Yet, despite the inclusion of ice, it is acknowledged that a simple bulk microphysics parameterization does not eliminate uncertainties in cold pool strength.

The model domain is 80 km x 80 km x 20 km with a horizontal grid spacing of 500 m. The vertical grid is stretched; the vertical grid spacing is 50 m below 1 km (the lowest grid level was at 25 m) and increases to 500 m above 12 km. Grid stretching allows for increased computational efficiency without sacrificing resolution at low levels, where it is important to resolve the imposed shallow stable layers. Open, wave-radiating boundary conditions are employed at the lateral boundaries following Klemp and Wilhelmson (1978b); the upper and lower boundaries are rigid and free-slip. Beneath the upper boundary a Rayleigh-damping sponge layer is applied above 14 km. As the model integrates through time, the grid is translated in order to keep the storm near the center of the domain.
2.2 Initialization

2.2.1 Parameter Space of Environmental Thermodynamic Profiles

The control simulation is initialized with a horizontally homogeneous environment using the analytic sounding of Weisman and Klemp (1982) in which the environmental potential temperature ($\tilde{\theta}$) and relative humidity ($\tilde{H}$) are determined by

$$\tilde{\theta}(z) = \begin{cases} \theta_o + (\theta_{tr} - \theta_o) \left( \frac{z}{\theta_{tr}} \right)^{5/4}, & z \leq \theta_{tr} \\ \theta_{tr} \exp \left[ \frac{g}{c_pT_{tr}} (z - \theta_{tr}) \right], & z > \theta_{tr}, \end{cases}$$

(2.1)

and

$$\tilde{H} = \begin{cases} 1 - \frac{3}{4} \left( \frac{z}{\theta_{tr}} \right)^{5/4}, & z \leq \theta_{tr} \\ 0.25, & z > \theta_{tr}, \end{cases}$$

(2.2)

where the tropopause height $\theta_{tr}$ is 12 km, the tropopause potential temperature $\theta_{tr}$ is 343 K and the surface potential temperature $\theta_o$ is 300 K. The water vapor mixing ratio at the surface is 14 g kg$^{-1}$. This sounding has surface-based CAPE of 2096 J kg$^{-1}$, surface-based CIN of 41 J kg$^{-1}$, and a lifting condensation level of approximately 1200 m. All CAPE and CIN calculations referenced in this study include the effects of moisture on buoyancy, neglect freezing, and are derived from the pseudoadiabatic ascent of a parcel lifted from the surface.

In each subsequent experiment, this temperature profile was modified by the addition of a stable surface layer (Fig. 2.1) with no change to the sounding above the stable layer. The imposed stable layers have a constant lapse rate and $\tilde{\theta}_{stb}(z)$ profile that depends on the amplitude of the surface temperature deficit relative to the original sounding ($\tilde{\theta}_o - \tilde{\theta}_{o, stb}$) and stable layer depth $\theta_{stb}$, that is,

$$\tilde{\theta}_{stb}(z) = \tilde{\theta} + \left[ \frac{\tilde{\theta}_o - \tilde{\theta}_{o, stb}}{\theta_{stb}} \right] (z - \theta_{stb}).$$

(2.3)

The stable layer depth $\theta_{stb}$ in each experiment ranges from 100 m to 1 km and the surface temperature deficit relative to the control sounding varies from 2.5 K to 10 K (stronger surface temperature deficits were used in several simulations.
Figure 2.1. The analytic sounding (Weisman and Klemp 1982) used in each simulation. Mixing ratio (g kg\(^{-1}\)) is shown in green, whereas temperature (°C) in the control simulation is shown in red. Near the surface, an example stable boundary layer (500 m deep with a temperature deficit of 5°C is shown in blue. The characteristics of this inversion are altered in each elevated simulation.

not shown here). There is no change to the water vapor mixing ratio profile (though low-level cooling alters relative humidity) in the experimental soundings except in cases where the introduction of a stable layer results in supersaturation. In these cases, the low-level water vapor mixing ratio is reduced such that the environmental relative humidity is 95%. It should be noted that the effects of
Figure 2.2. A matrix showing the stable boundary layer used in each experiment arranged by varying depth and surface temperature deficit. Temperature is in red, while mixing ratio profile (green) is consistent with the sounding in Fig. 2.1, except in cases of supersaturation. Above the displayed height, each sounding is identical to that in Fig. 2.1.

changing the low-level relative humidity may also affect the potential for tornado development. However, nocturnal boundary layers are associated with increases in relative humidity as the surface radiatively cools. Because these simulations most closely resemble such a boundary layer, some degree of control in the experiment
The changes near the surface to the temperature (and in some cases, moisture) profile in each experiment are depicted in Fig. 2.2, and the resulting thermodynamic characteristics are summarized in Table 2.1. The amount of CIN increases (decreases) as the amplitude of the surface temperature deficit increases (decreases) for a given stable layer depth. It also often increases (decreases) as the depth of the stable layer increases (decreases). However, in some cases, the surface temperature deficit is small while the stable layer is deep. As such, the lapse rate is not as stable as in those simulations having the same surface temperature deficit yet shallower boundary layers. Accordingly, the 1000 m deep stable layers with surface temperature deficits of 2.5 K and 5 K actually have less low-level CIN than their 500 m deep counterparts. Thus, CIN in each simulation is dependent not only on the stable layer depth or surface temperature deficit amplitude individually, but on a combination of both factors. The implementation of a stable boundary layer also results in changes to surface-based CAPE; however, the CAPE of the most unstable parcel in each profile remains relatively consistent with the control.
experiment at approximately 2450 J kg\(^{-1}\). This parcel is located above the stable boundary layer in each profile.

The downdraft convective available potential energy (DCAPE) of a parcel originating at 2 km also is computed for each profile. Unlike CAPE or CIN, which are conventionally defined for either a surface-based parcel, the most-unstable parcel, or a mixed-layer, DCAPE is defined using parcels from the level of origin of a downdraft. Considering that downdraft origin is difficult to determine and often varies considerably, a constant downdraft origin elevation is chosen that is consistent with downdraft trajectory calculations in this study. Furthermore, any change in DCAPE between experiments is due to the imposed stable layer below 2 km. Thus, the measure of DCAPE selected here is meant only as a qualitative indicator of the expected effect of the stable boundary layer on downdraft intensity rather than a precise predictor of downdraft speed.

### 2.2.2 Environmental Wind Profile

A clockwise-turning hodograph (Fig. 2.3) similar to those used by both Rotunno and Klemp (1982; their Fig. 5) and Weisman and Klemp (1986; their Fig. 17) is used in all of the experiments, though straight and semicircular hodographs were used in sensitivity tests. In reality, low-level shear often changes dramatically in the presence of a stable-boundary layer. However, in order to isolate the effects of only low-level stability on each simulation the same wind profile was used in all simulations within and above the stable boundary layer. In this hodograph, the wind shear vector turns anticyclonically (i.e. veers) with height from the surface to 2 km and is unidirectional from 2 km to 6 km. Above 6 km, the wind is constant with height. Domain translation is implemented by subtracting the assumed storm motion from the winds at each level, effectively making the storm stationary within the domain. Because CM1 produces grid-relative wind fields, all subsequent horizontal wind field results should be considered in this context relative to storm features (e.g. updrafts). The chosen hodograph results in the development of splitting storms in each simulation; however, attention is primarily focused on the dominant right-moving (cyclonically rotating) supercell.
2.2.3 Mechanism for Convection Initiation

All of the storms are triggered using the method of Klemp and Wilhelmson (1978b). A positively buoyant bubble with a maximum potential temperature perturbation of 2 K is inserted near the center of the horizontal domain at the lower vertical grid levels. The perturbation bubble has a horizontal radius of 10 km, a vertical radius of 1.5 km and is initially centered at 1.5 km above ground level. Thus, in every experiment the center of the bubble is located above the imposed stable layer. This initiation technique is physically unrealistic, but it offers the advantage of quickly promoting convection that rapidly evolves into realistic storm structures. Consequently, after the initial storm spin-up there is little to no trace of the triggering mechanism to contaminate the results. The greatest advantage of initializing a storm with this method is that the bubble, obviates the need to introduce some other horizontal heterogeneity into the environment. Although this would be more realistic (after all, most storms form along convergent air mass boundaries, and

Figure 2.3. The idealized, clockwise-turning hodograph used in the initial vertical wind profile for each simulation with heights labeled every km AGL. The averaged storm motion vector (red arrow) is also plotted.
the boundaries are associated with gradients in wind and usually temperature and moisture), the introduction of horizontal heterogeneity unavoidably makes comparisons of the results from case to case more difficult, because differences in evolution will be due to horizontal heterogeneity in addition to differences in the mean environmental conditions (Richardson et al. 2007).
Chapter 3

Results

In the forthcoming presentation of the simulation results, the control supercell is compared with those situated over a stable boundary layer at 85 min (5100 s) in each simulation unless otherwise noted. This corresponds to the time at which the control supercell displays its maximum near-surface (nominally 25 m, which is the lowest grid level for horizontal winds) vertical vorticity. The risk in performing comparisons at a common time is that storms may evolve at different rates, such that differences among storms observed at a common time may not reflect fundamental dynamical differences among the storms as much as simply differences in the time at which certain aspects of the storm develop [e.g., there is some risk that an elevated storm differing considerably from the control storm at 85 min could look very similar to the control storm (at 85 min) 10-20 minutes later]. Thus, there might be advantages to comparing storms at a common evolutionary milestone instead, e.g., the time of maximum vertical vorticity at some level. The downside of this latter approach, however, is that the choice of such a milestone is subjective by nature and might mask significant dynamic differences. For instance, if every storm were compared at the time of its maximum low-level vertical vorticity they might all appear at that instant to be more similar to the control simulation than is really the case over the rest of their lifetime.

In this investigation, simulated storms were compared both at common times as well as common evolutionary milestones (time of maximum near-surface vertical vorticity). It was found that, whether the comparisons were made at a common time or common evolutionary milestone, the same conclusions can be drawn from
the results of the numerical simulations. Given the choice between two relatively equitable options, the former method of comparison was selected because it is the most objective in that each storm has had the same amount of time to develop and the arbitrary choice of a milestone is unnecessary. As such, the differentiating factor in each comparison remains the storm environment, not the simulation time.

### 3.1 General Relationship Between Simulated Supercell Characteristics and the Low-Level Temperature Profile

In each simulation, the initial warm bubble results in organized convection that can be classified as a supercell thunderstorm (or at least a thunderstorm with transient supercell characteristics in the case of the 1km10c simulation). By 30 minutes into the simulation, the storms attain quasi-steady vertical velocities (Fig. 3.1), both at midlevels and near the surface, and quasi-steady vertical vorticity at midlevels (Fig. 3.2b). Significant rotation develops near the surface by 60 minutes, at least within the storms that develop significant near-surface rotation (Fig. 3.2a). Table 3.1 summarizes the maximum vertical vorticity values attained in each simulation, as well as the maximum and minimum vertical velocities attained.

An examination of the control simulation at 85 min at low levels (Fig. 3.3) and midlevels (Fig. 3.4) reveals a mature right-moving supercell with traits similar to those of the conceptual model of Lemon and Doswell (1979). Although the horizontal resolution is too coarse to resolve a tornado explicitly, the control supercell exhibits strong rotation near the ground. At the lowest grid level, the storm attains a maximum in vertical vorticity of $6.3 \times 10^{-2} \text{s}^{-1}$. This vorticity maximum is co-located with an area of rising motion that extends southward away from the precipitation region, along and ahead of the gust front. The control storm also displays a low-level downdraft region that wraps around the low-level mesocyclone on its western flank. A strong midlevel updraft (e.g. $28 \text{ m s}^{-1}$ at 4 km AGL; Fig. 3.3) is located above the near-surface vertical vorticity maximum. The midlevel vorticity maximum lies more toward the rear flank of the storm, yet is still within updraft.
Figure 3.1. Time series of maximum vertical velocity (updraft speed) at low-levels (dotted dashed lines) in the right-moving supercell for the control (red), 500m5c (dashed blue) and 1km10c (dotted dashed black) simulations.

The right panels in Figs. 3.3 and 3.4 display a matrix of horizontal cross-sections depicting the salient traits of the nine supercells simulated in environments having a low-level stable layer (cf. Fig. 2.2) at the same time and heights as the control supercell. A first-glance inspection of these cross-sections reveals storms trending away from the classic tornadic supercell model as the depth and amplitude of the stable layer increase. The simulations to the lower-left (i.e., those initialized with
Figure 3.2. Time series of maximum vertical vorticity at (a) low-levels and (b) midlevels in the right-moving supercell for the control (red), 500m5c (dashed blue) and 1km10c (dotted dashed black) simulations.

Table 3.1. Summary of relevant model results for each simulation.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Surface-based CAPE (J kg⁻¹)</th>
<th>Surface-based CIN (J kg⁻¹)</th>
<th>2 km DCape (J kg⁻¹)</th>
<th>Low-level ᴱμᵥ max (10⁻⁵ s⁻¹)</th>
<th>Midlevel ᴱμᵥ max (10⁻⁵ s⁻¹)</th>
<th>Low-level w max (m s⁻¹)</th>
<th>Low-level w min (m s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>control</td>
<td>2096</td>
<td>41</td>
<td>254</td>
<td>6.27</td>
<td>6.41</td>
<td>5.73</td>
<td>-10.61</td>
</tr>
<tr>
<td>100m2.5c</td>
<td>1811</td>
<td>84</td>
<td>253</td>
<td>5.66</td>
<td>5.96</td>
<td>5.00</td>
<td>-11.19</td>
</tr>
<tr>
<td>100m5c</td>
<td>1548</td>
<td>128</td>
<td>253</td>
<td>5.79</td>
<td>5.64</td>
<td>6.21</td>
<td>-9.62</td>
</tr>
<tr>
<td>100m10c</td>
<td>1065</td>
<td>210</td>
<td>251</td>
<td>4.73</td>
<td>4.24</td>
<td>5.16</td>
<td>-8.60</td>
</tr>
<tr>
<td>500m2.5c</td>
<td>1601</td>
<td>104</td>
<td>241</td>
<td>5.43</td>
<td>4.67</td>
<td>5.71</td>
<td>-8.00</td>
</tr>
<tr>
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<td>1160</td>
<td>164</td>
<td>225</td>
<td>4.44</td>
<td>4.41</td>
<td>4.21</td>
<td>-8.09</td>
</tr>
<tr>
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<td>6</td>
<td>562</td>
<td>195</td>
<td>2.47</td>
<td>4.76</td>
<td>4.02</td>
<td>-7.27</td>
</tr>
<tr>
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<td>88</td>
<td>221</td>
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<td>4.79</td>
<td>5.67</td>
<td>-8.16</td>
</tr>
<tr>
<td>1km5c</td>
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<td>184</td>
<td>6.42</td>
<td>4.33</td>
<td>4.93</td>
<td>-7.99</td>
</tr>
<tr>
<td>1km10c</td>
<td>NO LFC</td>
<td>NO LFC</td>
<td>125</td>
<td>2.65</td>
<td>4.41</td>
<td>3.28</td>
<td>-6.76</td>
</tr>
</tbody>
</table>

only a shallow, small-amplitude stable layer) are visually similar to the control in that they display hook-like structures in their precipitation fields, prominent low-level vorticity maxima, and well-defined updrafts along their gust fronts. In contrast, these features are less distinguishable or even absent in the upper-right simulations (i.e., those initialized with a deep, large-amplitude stable layer).
Figure 3.3. Low-level, horizontal cross-sections of the control (left) and stable-layer (right) simulations at 5100 s (85 minutes). Stable-layer simulations are organized with depth as the ordinate and surface temperature deficit as the abscissa. Green shaded regions represent rainwater mixing ratio $>$ 1 g kg$^{-1}$ at 125 m AGL. Updraft (downdraft) velocity at 125 m AGL is contoured in solid (dashed) black at intervals of 1 m s$^{-1}$. Vertical vorticity is contoured in red at 25 m AGL in 0.01 s$^{-1}$ intervals. Maximum values of vertical vorticity and updraft/downdraft speed within the horizontal domain are reported in the lower right corner of each cross-section.

3.1.1 Relationship between Storm Characteristics and CIN

CIN is compared with the vertical vorticity and vertical velocity maxima at midlevels and near the surface (Fig. 3.3 and Table 3.1). It should be noted that additional simulations beyond those presented in Figs. 3.3 and 3.4 are included in calculations of correlations in order to provide a larger sample size. These simulations are initiated with stable layer depths and surface temperature deficits within the range of those described in Table 2.1. Simulations in which there is no (surface-based) LFC are not included in the correlation calculations (CIN is undefined in such environments). Furthermore, correlation coefficients were calculated both including and excluding two simulations with relatively high CIN that are considered outliers. In the following discussion of results, correlations excluding these outliers will be presented.

The results displayed in Fig. 3.5a illustrate a negative correlation, with a cor-
Figure 3.4. Midlevel, horizontal cross-sections of the control (left) and stable-layer (right) simulations at 5100 s (85 minutes). Same is in Fig. 3.3 but vertical vorticity and vertical velocity cross-sections are 4 km AGL. Vertical velocity contours are at 5 m s^{-1} intervals.

relation coefficient of -0.79 between CIN and low-level maximum vertical vorticity. This general relationship is also evident for 85 min vertical vorticity maxima (Fig. 3.3). At that time the near-surface vertical vorticity decreases by 93% between the control mesocyclone and the simulation with the strongest CIN and weakest low-level rotation (1km10c). At 4 km AGL, CIN and maximum vertical vorticity are less strongly anti-correlated (Fig. 3.5b, correlation coefficient of -0.55; it should be noted that when outliers are considered there is not a significant correlation between CIN and midlevel rotation). At 85 min, however, the simulations with the largest CIN do not exhibit the weakest rotation. For example, the simulation with the strongest midlevel vertical vorticity (1km10c) besides the control has no surface-based LFC. This suggests that though midlevel rotation does decrease with increased CIN, it is less sensitive to changes in CIN than low-level rotation.

Maximum updraft strength generally decreases at low-levels (Fig. 3.5c) as CIN increases. At low levels, the vertical velocity maximum at 85 min decreases by nearly 60% from the control updraft to the simulation with the weakest updraft (1km10c). At midlevels, the updraft speed maxima over all times display a similar negative correlation with CIN to those at low levels. However, at 85 min the
Figure 3.5. Graphs of CIN vs. maximum (a) low-level vertical vorticity, (b) midlevel vertical vorticity, (c) low-level vertical velocity, and (d) midlevel vertical velocity for a range of varying stable-layer simulations (some simulations are included that are not shown in Figs. 3.3 and 3.4). All maxima are computed over the entire horizontal domain at specified level over all times. A linear trend line and correlation coefficient are also included. Those simulations marked with a green asterisk are considered outliers due to their anomalously high CIN. Correlation coefficients computed without them are in parentheses.

Midlevel updraft maxima do not show any significant variation with CIN. In fact, the control simulation produces a weaker updraft (second only to the 1km10c
storm) than most of the simulations containing much greater CIN. With only a 15% decrease in updraft magnitude between the strongest and weakest case, the presence of a stable layer only marginally affects the midlevel updraft.

### 3.1.2 Relationship between Storm Characteristics and DCAPE

Whereas CIN is an adequate predictor of low-level updraft intensity, DCAPE is more relevant in assessing a stable layer’s influence on low-level downdraft strength. As such, DCAPE is computed for a downdraft parcel originating at 2 km in the base state sounding of each simulation. As both the depth and surface temperature deficit of the stable layer increase, DCAPE decreases from a control value of 254 J kg\(^{-1}\) to 125 J kg\(^{-1}\) for the 1km10c simulation. The simulation results show that a relationship exists between DCAPE and low-level downdraft magnitude as well as DCAPE and vertical vorticity. In particular, DCAPE of parcels originating 2 km AGL is positively correlated (coefficient of 0.81) with maximum low-level downdraft magnitude (Fig. 3.6a). DCAPE also shows a moderate positive correlation to both low and midlevel vertical vorticity (Figs. 3.6b and 3.6c, respectively). In summary, decreased DCAPE is associated with weaker low-level downdrafts and weaker vertical vorticity.

![Graphs of DCAPE for a downdraft parcel originating 2 km AGL vs. maximum (a) low-level downdraft magnitude, (b) low-level vertical vorticity, and (c) midlevel vertical vorticity. All maxima are computed over the entire horizontal domain at the specified level over all times.](image-url)

**Figure 3.6.** Graphs of DCAPE for a downdraft parcel originating 2 km AGL vs. maximum (a) low-level downdraft magnitude, (b) low-level vertical vorticity, and (c) midlevel vertical vorticity. All maxima are computed over the entire horizontal domain at the specified level over all times.
3.1.3 Relationship between Maximum Updraft/Downdraft at Low Levels and Maximum Vertical Vorticity at Low Levels

Given the findings in the literature that emphasize the importance of both the updraft and downdraft in the development of vorticity near the surface (Rotunno and Klemp 1985; Davies-Jones and Brooks 1993), one might expect some relationship between maximum vertical velocity and maximum vertical vorticity at low levels. Besides being co-located with updraft at low levels (Fig. 3.3), maximum vertical vorticity also tends to increase with increasing maximum vertical velocity across the parameter space of simulations. Fig. 3.7a shows a positive correlation (coefficient of 0.68) but significant spread between maximum vertical velocity and vertical vorticity near the ground. Furthermore, the control (1km10c) simulation exhibits both the largest (smallest) low-level vorticity and largest (smallest) low-level vertical velocity. Intermediate simulations show a similar relationship. A similar, though slightly weaker, relationship exists between low-level downdraft magnitude and low-level vertical vorticity (Fig. 3.7b). With a correlation coefficient of 0.52, maximum downdraft strength is somewhat correlated to maximum low-level rotation.

It should be made clear that in the foregoing correlation analysis, one cannot infer causality. Though relationships do exist between vertical velocity and vertical vorticity, correlation alone cannot determine if a weaker downdraft or updraft causes decreased rotation in storms over stable boundary layers. It could be equally likely (given this evidence) that stronger vertical vorticity generates larger vertical velocities at low levels, or that both vorticity and velocity are dependent on an outside process that varies between simulations. What is evident, however, is that a general relationship exists between vertical velocity (both positive and negative) and vertical vorticity at low levels. The causes of this relationship are further explored in Chapter 4.
3.2 Trajectory Analysis of Airflow within Elevated Supercells

Under the definition of elevated convection adopted in this study, the updraft of a simulated supercell must not be able to draw inflow from near the surface. In the presence of a stable boundary layer, one might also expect that downdraft parcels become less negatively buoyant and may not be capable of reaching the ground. As a means of determining which stable boundary layer environments result in elevated convection and how vertical airflow is altered in such storms, trajectories of updraft and downdraft parcels are computed. Three-dimensional trajectories are calculated from the model output wind fields using a fourth-order Runge-Kutta scheme. The velocity field is spatially interpolated to parcels from the model grids every 15 s (the timestep at which model data is saved).

Figure 3.7. Scatter plot of (a) low-level maximum vertical velocity vs. low-level maximum vertical vorticity and (b) low-level maximum downdraft magnitude vs. low-level maximum vertical vorticity. All maxima are computed over the entire horizontal domain at the specified level over all times.
3.2.1 Updraft Parcel Origins

The origins of midlevel updraft parcels are determined during the mature phase of each simulated storm using back trajectories. Updraft parcel trajectories are initiated from 85 min and calculated backwards 15 minutes. Grid points with vertical velocities greater than 2 m s\(^{-1}\) at 4 km AGL within a 400 km\(^2\) subset of the horizontal domain containing the storm of interest (the selected sub-domain of each simulation is the same as shown in Fig. 3.3) are chosen as updraft parcels. The origin of each parcel is defined in this analysis as simply the height of each updraft trajectory at the final integration time step.

![Image](image.jpg)

**Figure 3.8.** (a) horizontal cross-section at 4 km of parcels with vertical velocity > 2 m s\(^{-1}\) colored according to height of parcel origin (m). (b) Back-trajectories of updraft parcels lying along line segment AB in (a).

Trajectory analysis shows that the control supercell is clearly surface-based in that some of its updraft parcels have origins near the surface. Fig. 3.8a presents the spatial distribution of the origin of updraft parcels. Parcels nearest the western flank of the updraft have the lowest origins with a gradual elevation in parcel origin height towards the eastern flank of the updraft. The lowest parcel ascends from 32 m AGL whereas some parcels have sources several km AGL or even above the initial height of the updraft parcel. A normalized histogram classifying updraft parcels by origin height (Fig. 3.9) reveals that a minority of updraft parcels originate in the
Figure 3.9. Normalized histogram of 4 km updraft parcel origin altitude (km AGL) for control simulation. The horizontal line represents height of updraft parcel destination.

lowest height bin. One might find it surprising that a large majority of the updraft parcels come from above the surface layer in an environment where there is little to inhibit surface parcels from ascending. However, the existence of surface-based CIN, even in the control case, results in the most unstable parcel in the base state sounding being above the surface layer. Thus, potential positive buoyancy and vertical accelerations are not limited to only surface parcels. Furthermore, some of these parcels are likely from the midlevel environment and have been entrained into the initial surface-based updraft. This conclusion is supported by Fig. 3.8b which clearly shows a horizontal component to many of the updraft trajectories that is indicative of entrainment.

Analysis of the source of midlevel updraft trajectories within stable boundary layer simulations yields the striking result that many of the storms in these cases
are not truly elevated. The spatial pattern of updraft parcel origin in each stable layer case is largely similar to that of the control supercell. Normalized histograms of updraft parcels sorted by source height are presented in Fig. 3.10. In many of the stable boundary layer simulations, at least some updraft parcels have origins within the lowest grid levels. All of the simulations with a 100 m deep stable boundary layer remain surface-based. In three of the six cases with deeper stable layers, the lowest bin containing parcels is below 100 m. Following the adopted
definition of elevated convection, only the 500m10c, 1km5c, and 1km10c storms can be confidently classified as elevated. However, even in these cases, the storms are capable of drawing inflow from within at least the top of the stable boundary layer.

### 3.2.2 Downdraft Parcel Destinations

The destination altitudes of 2 km AGL downdraft parcels are determined for each simulation using trajectories run forward for 10 minutes. The trajectories are initiated at 85 min for parcels having a vertical velocity less than -2 m s\(^{-1}\) in a 400 km\(^2\) horizontal subset of the domain encompassing the storm of interest. The destination altitude of each parcel is simply its height at the last trajectory time step.

![Figure 3.11](image)

**Figure 3.11.** Similar to Fig. 3.9, but for destinations of downdraft parcels originating at 2 km AGL.
Analysis of normalized histograms of the control (Fig. 3.11) and stable boundary layer (Fig. 3.12) simulations shows that numerous downdraft parcels are capable of reaching the surface despite the presence of a strongly stable boundary layer. As one might expect, in the absence of a stable boundary layer nearly twice as many downdraft parcels end up in the lowest height bin than at any other level. Less intuitively, in over half of the stable layer simulations the lowest height bin contains more parcels than any other single level. In the remaining stable layer cases, a significant number of downdraft parcels are still able to penetrate the
stable boundary layer to the lowest bin level. At first glance it may also seem surprising that some downdraft parcels have destinations that are well above their origin (a few reach as high as 13 km AGL in some simulations). However, this is easily explained by entrainment of downdraft parcels into the updraft well above the ground or the recycling of downdraft parcels at the surface into the updraft.

It appears that there is a relationship between DCAPE and parcel destination. The simulations in which the lowest level is the preferred destination for downdraft parcels (control, 100m2.5c, 100m5c, 100m10c, 500m2.5c) also have the highest and similar values of 2 km DCAPE (all have approximately 250 J kg$^{-1}$ of 2 km DCAPE). Alternatively, those parcels with the weakest values of 2 km DCAPE (500m5c, 500m10c, 1km2.5c, 1km5c, 1km10c) tend to have fewer downdraft parcels reach the ground.

3.3 Comparisons of the Control Supercell with Extreme and Intermediate Stable Boundary Layer Cases

The dynamics responsible for the suppression of near-surface rotation in the elevated supercells are best understood through a more detailed comparison of the control simulation with two simulations that have less low-level vertical vorticity. Below is a comparison of the differences between the control supercell and an elevated supercell that develops within the environment initialized with a 1 km-deep stable layer with a surface temperature deficit of 10°C. The control is also compared against the marginally elevated 500m5c supercell in which the dynamical differences are more subtle.

The 500m5c simulation is chosen for comparison with the control simulation because of its realistic depth and surface temperature deficit relative to the parameter space investigated. These characteristics are consistent with observations by Stull (1983) as well as Nieuwstadt and Driedonks (1979) of nocturnal boundary layer height and surface temperature deficit. Both studies observed nocturnal stable boundary layers with heights on the order of several hundred meters. Stull (1983) found surface temperature deficits of anywhere from 4 K to 13 K whereas
Nieuwstadt and Driedonks (1979) observed a mature nocturnal boundary layer in which the surface temperature decreased by nearly 5 K. Therefore, the 500 m deep stable layer case with a 5 K surface temperature deficit is adequately representative of a fully developed nocturnal boundary layer. Although it may be a less realistic representation of a nocturnal boundary layer, the 1km10c case is also chosen because it has been shown to be fully elevated. Consequently, the differences that exist between the control and 500m5c simulations may be clarified by comparison with this more extreme case in which the updraft is entirely decoupled from the surface. The CIN of the environmental soundings in both the 500m5c (164 J kg\(^{-1}\)) and 1km10c (infinite CIN due to the lack of an LFC) is considerably stronger than the CIN of the control sounding (41 J kg\(^{-1}\)).

![Figure 3.13](image)

**Figure 3.13.** Time series of maximum downdraft magnitude at (a) low-levels and (b) midlevels in the right-moving supercell for the control (red), 500m5c (dashed blue) and 1km10c (dotted dashed black) simulations.

Maximum vertical vorticity time series (Fig. 3.2) reveal that rotation is nearly always stronger in the control supercell near the surface, whereas at midlevels the difference between the control and stable boundary layer simulation mesocyclones is not as evident. The control supercell rapidly develops near-surface rotation around 30 min (1800 s), with maximum low-level vertical vorticity of 0.063 s\(^{-1}\) occurring at 85 min (5100 s). The 500m5c and 1km10c storms spin up low-level
rotation at approximately the same time, but generally have less intense vertical vorticity maxima than in the control supercell. Whereas the 500m5c simulation is at least capable of generating vertical vorticity above $0.02 \text{ s}^{-1}$, the 1km10c simulation struggles to attain values above $0.01 \text{ s}^{-1}$. On the other hand, midlevel vertical vorticity is similar in all three cases through time (Fig. 3.2b), with the exception of after 5400 s when the control has generally stronger midlevel rotation. These results are in agreement with the finding in Section 3.1.1 that near-surface vertical vorticity decreases as CIN increases, and midlevel vertical vorticity is less sensitive to changes in CIN.

Time series of maximum vertical velocity (Fig. 3.1) suggest relationships between CIN and updraft strength at low and midlevels similar to those presented in Section 3.1.1. At low levels the control updraft is generally stronger than both the 500m5c and 1km10c simulations, with the 1km10c simulation typically exhibiting the weakest updraft through time. The 1km10c storm also has the weakest midlevel updraft. Though the control supercell attains a higher overall maximum vertical velocity than the 500m5c storm, they both fluctuate over time around a similar updraft speed. Both vertical vorticity and vertical velocity time series indicate the observed relationship between storm strength at low levels and CIN is consistent over the duration of the simulation, not only at the moment analyzed in Figs. 3.3 and 3.4 (85 min). This implies that the results discussed in Section 3.1.1 are robust across all times after the initial storm development.

On average, the low-level downdraughts are stronger in the control supercell than in the elevated and marginally elevated storms, especially prior to 4800 s. At midlevels, however, the average downdraughts are of comparable strength between the control and 500m5c storms, with the 1km10c simulation exhibiting a slightly weaker midlevel downdraft at most times (Fig. 3.13). Additionally, there is considerably more variability in the low-level downdraft magnitude over time than is observed in the updraft magnitude. These results are not unexpected given the dependence of low-level downdraft magnitude on DCAPE discussed in Section 3.1.2. Although the near-surface downdraft is weaker in the two stable layer cases, the existence of a downdraft at low-levels across all times suggests that there is no time at which the downdraft is unable to penetrate even the most stable boundary layer.
Chapter 4

Analysis and Discussion

4.1 Storm Airflow

4.1.1 Updraft Origins

The results presented in Section 3.2.1 indicate that in many of the simulations including a stably-stratified boundary layer storms are capable of processing air from within the stable layer if not from the surface. How then do these negatively buoyant parcels overcome significant CIN and acquire positive vertical velocity? The most likely explanation is a dynamic vertical pressure gradient induced by the rotating midlevel updraft and veering shear profile. In the control simulation, positive vertical vorticity is associated with the area of the midlevel updraft where parcels have their lowest source (Figs. 3.4 and 3.8). A similar spatial relationship exists for the stable boundary layer supercell updrafts as well. This suggests that the midlevel mesocyclone generates a dynamic vertical perturbation pressure gradient favorable for lifting negatively buoyant low-level parcels to their LFC. The veering of the shear vector with height characteristic of the environmental wind profile also creates an upward-directed dynamic vertical perturbation pressure gradient force in these right-moving storms (cf. Section 1.1).

Corfidi et al. (2008) concluded that the binary classification of convection as either elevated or surface-based overextends our current understanding of severe storm processes. They argue for a “continuum of convection” in which convection ranges from surface-based to elevated with intermediate shades of gray or “de-
degrees of elevation”. Based on the findings in this study, the author supports this viewpoint. The elevated characteristics of a storm over a surface boundary layer appear to be tied to more than simply the depth or surface temperature deficit of the stable layer. Because midlevel updraft rotation, and therefore dynamic vertical pressure gradients, is more likely dependent on environmental wind shear than on low-level thermodynamic characteristics, CIN alone cannot determine how elevated a storm will be. More generally, storms initiated in similar thermodynamic profiles may not always exhibit the same “degree of elevation”.

4.1.2 Downdraft Destinations

The results presented in Section 3.2.2 show that downdraft parcels are capable of reaching the surface despite the existence of a strong stable boundary layer. Although one may expect a strongly stable stratification to prevent a downdraft from penetrating to the surface by eliminating the negative buoyancy of a downdraft parcel, even the simulation with the most stable boundary layer exhibits downdraft parcels that reach the surface. It is important to note that every simulation has DCAPE for parcels originating at 2 km AGL and therefore, every 2 km downdraft experiences a net negative buoyancy force throughout its descent. Despite this, some simulations (those with a 10°C surface deficit) result in 2 km origin downdrafts that may actually become positively buoyant at very low levels. This is because in these simulations the downdraft parcel becomes warmer than the environmental temperature as it descends, yielding a small area of “downdraft inhibition” that acts to decelerate the downdraft. Given the presence of DCAPE despite significant surface static stability, it becomes clear why downdraft parcels are capable of reaching the surface even in the most stable of the simulations. Even in cases where the downdraft eventually experiences deceleration, at least some downdraft parcels have enough downward momentum to reach the surface because the depth over which they experience an upward acceleration is very shallow. In order to model a downdraft that is incapable of reaching the surface, the simulation would require an environmental sounding where the area of downdraft inhibition was much deeper than even in the 1km10c case presented here.

This is not to say that the influence of a stable boundary layer on a storm’s
downdraft is insignificant in regards to producing low-level rotation. Though downdrafts may still reach the surface through stable boundary layers, they are certainly weakened at low levels relative to a sounding with little or no static stability at the surface (Fig. 3.13a). Conceivably, this weakening could possibly lead to less generation of horizontal vorticity by the downdraft that might aid subsequent vertical vorticity generation at the surface. Furthermore, weakened downdrafts may decrease the tilting of horizontal vorticity into a vertical orientation. This conclusion will be further discussed in the following sections.

4.2 Circulation Analysis

In order to elucidate the dynamical reasons why the control supercell contains stronger near-surface vertical vorticity than the marginally elevated and elevated supercells, an examination of the evolution of material circuits encircling the low-level vertical vorticity maxima (and the circulation about those circuits) is useful. The method described below follows the examples of Rotunno and Klemp (1985) and Davies-Jones and Brooks (1993).

Circulation, $C$, is defined as:

$$ C \equiv \oint \vec{v} \cdot d\vec{l} \tag{4.1} $$

where $\vec{v}$ is the velocity vector and $\vec{l}$ is the position vector of a point on the material circuit that the line integral is calculated around. Under the inviscid, Boussinesq approximation, the following time tendency of circulation, known as Bjerknes circulation theorem, may be derived (Rotunno and Klemp, 1985):

$$ \frac{dC}{dt} = \oint B\vec{k} \cdot d\vec{l} = \oint Bdz \tag{4.2} $$

where $B$ is the buoyancy and $\vec{k}$ is the vertical unit vector. Equation 4.2 implies that circulation is generated (destroyed) around a material circuit as the height of the circuit decreases (increases) in an area of negative buoyancy.

Stokes theorem may be used to show that the average vorticity normal to the area enclosed by a material circuit is simply the circulation around the circuit.
divided by the area within the circuit. Mathematically, this can be shown as:

\[ C = \oint \vec{v} \cdot d\vec{l} = \iint_{A} (\nabla \times \vec{v}) \cdot \vec{n} dA \]  

(4.3)

where \( A \) is the area bounded by the material circuit and \( \vec{n} \) is the unit vector normal to that area at all places. Thus, for a horizontal circuit, the circulation around it is proportional to the average vertical vorticity contained therein.

### 4.2.1 Circulation around Instantaneous Vorticity Maxima

In order to determine how much vertical vorticity is available to be stretched by a storm’s updraft throughout the life cycle of the simulations, the circulation around the low-level vorticity maximum is calculated for the control, 500m5c, and 1km10c supercells. The circulation is computed around a horizontal, annular circuit with a radius of 2 km situated at 50 m AGL and centered around the low-level vorticity maximum at every time step. Because the circuit is horizontal, Equation 4.3 can be used to estimate the circulation to a good approximation by multiplying the average vertical vorticity within the circuit by the area it encloses.

The results of these calculations are presented in the time series shown in Fig. 4.1. Not surprisingly, the evolution of circulation around each vorticity maximum is very similar to the evolution of low-level vertical vorticity for all three simulations. At nearly every time, the control exhibits stronger circulation than the 500m5c and 1km10c simulations, with the 1km10c being the weakest. Because the broad low-level circulation considerably weakens as low-level stability increases, it can be concluded that some other process in addition to decreased stretching by a weaker updraft contributes to less near-surface rotation in the stable boundary layer simulations. If only updraft stretching were responsible for the observed differences in low-level vertical vorticity, then one would expect the broad low-level circulation to be relatively constant between simulations. In other words, the amount of vertical vorticity over a broad area that has the potential to be converged and stretched by the updraft is unequal between the control and stable layer cases. Therefore a difference between simulations in the processes that lead to the low-level circulation must be at least partly responsible for decreased near-surface rotation in the elevated and marginally elevated cases.
Figure 4.1. Time series of circulation around 2 km radius fluid circuits centered on the low-level vorticity maximum at each time for the control (red), 500m5c (dashed blue) and 1km10c (dotted dashed black) simulations.

4.2.2 Evolution of Material Circuits Converging around Vorticity Maxima

To account for differences in circulation around the low-level vorticity maximum between simulations, the evolution of fluid circuits encircling the vorticity maxima at 81 minutes for the control, 500m5c, and 1km10c simulations is analyzed. This initialization time is selected because it allows the foregoing analysis to capture the development of the low-level circulation that is presumably converged into the observed vorticity maxima at 85 minutes (the time of maximum low-level vertical vorticity in the control simulation). The trajectories of a ring of 500 parcels are
computed 30 minutes backward in time from 81 minutes for each case. The circuits at 81 minutes have a radius of 2 km and an altitude of 75 m. By starting trajectories at this altitude with the relatively short integration timestep of 15 s, extrapolation of horizontal wind data to levels below the lowest grid level where they are computed by the model (25 m AGL) is largely avoided. Trajectories initiated below 75 m AGL have diminished credibility because significant extrapolation of horizontal wind data occurs in such cases. Final positions of each circuit relative to the 81 minute low-level vertical vorticity maximum and vertical velocity fields are shown in Fig. 4.2.

![Figure 4.2](image-url)

**Figure 4.2.** Final position of material circuit (red) for the (a) control, (b) 500m5c and (c) 1km10c simulations at 81 minutes. Letters on each circuit represent the initial position of reference parcels. Vertical vorticity at 75 m AGL is plotted in increments of 0.005 s\(^{-1}\) (blue). Vertical velocity at 125 m AGL is plotted in increments of 1 m s\(^{-1}\) (black, dashed contours are negative values). Arrows are horizontal wind vectors.

As the trajectories are calculated backwards through time (parcel locations are computed every 15 s) from the positions indicated in Fig. 4.2, the circulation is computed every 30 s in a counterclockwise fashion around each circuit. A time series of the circulation for each case is presented in Fig. 4.3. It is evident that the circulation around the fluid circuits varies significantly over the 30 minute time frame. Circulation around the control case circuit begins with significantly negative values, but grows to strongly positive values by 70 minutes. After a sharp drop in circulation to slightly negative values at approximately 73 minutes, circulation increases to a value of 6.8 \(\times 10^4\) m\(^2\) s\(^{-1}\). In the 500m5c simulation, circulation about the circuit begins near zero then becomes negative. By 57 minutes circulation...
is increasing to a maximum positive value near 68 minutes. From this point, circulation decreases gradually to a final value of $3.0 \times 10^4$ m$^2$ s$^{-1}$. Finally, around the 1km10c circuit, circulation begins slightly negative and remains as such until about 65 minutes when it becomes slightly positive before decreasing to a final value of $0.5 \times 10^4$ m$^2$ s$^{-1}$.

Figure 4.3. Time series of the circulation around a 2 km radius fluid circuit of 500 parcels centered around the 81 minute low-level vertical vorticity maximum that is followed 30 minutes backwards in time for the control (red), 500m5c (dashed blue) and 1km10c (dotted dashed black) simulations. The time at which the circuits are analyzed in Fig. 4.5 is marked by the vertical dotted line.

Similar circulation analysis were computed for these cases at other times. It is
important to note that although the evolution of circuits around low-level vorticity maxima may vary significantly depending on the time at which the circuit is initialized, in nearly all cases the control ends up with a larger circulation around its low-level vorticity maximum than in either of the stable-layer simulations. The 500m5c circuit is capable of accumulating similar values of circulation as the control; however, over the last several minutes of the circuits’ evolution, the control nearly always manages to generate more circulation than the 500m5c case. This suggests that a difference exists between the control and stable layer simulations late in the evolution of the circuits as they near the ground.

The general variability of circulation over time around these circuits is prominent and suggests that differences in the evolution of circulation, and thereby vertical vorticity, are not easily generalized between cases. Regardless, some information about the varying behavior of low-level vorticity may be gleaned from a close inspection of the material circuits as they converge around the 81 minute low-level vorticity maximum. It is clear from Fig. 4.3 that the control case generates significantly more circulation at low levels than either stable layer simulation. How does this occur? To answer this question, the geometry of each circuit as well as the circuit’s position relative to the low-level buoyancy field as it converges upon the low-level vorticity maximum must be explored further.

Fig. 4.4 illustrates the physical evolution of the circuits through space over the 15 minutes directly prior to their final position at 75 m AGL. In the control (Fig. 4.4a), the circuit initially has a complicated structure with nearly half of the parcels (those from C to A) originating from aloft at 66 minutes. These parcels descend (some from as high as 3 km) over the next 15 minutes to 75 m AGL at 81 minutes. This suggests that approximately half of the parcels participating in the low-level vorticity maximum had origins in the forward-flank downdraft. The rest of the parcels (those from A to C) have inflow origins. They originate well east of the storm and move westward with time (relative to the storm). By 76 minutes, the circuit has become considerably less complex as parcels approach the vorticity maximum.

The 500m5c circuit (Fig. 4.4b) displays a bifurcation similar to that in the control case, but with noticeable differences. In the 500m5c case (Fig. 4.4b), the circuit exhibits considerably less descent with time than in the control over the 15
Figure 4.4. Perspective views looking from the east-southeast of the material circuits for the (a) control, (b) 500m5c, and (c) 1km10c simulations at three 5-minute intervals prior to the final circuit position. Rainwater mixing ratio exceeding 1 g kg\(^{-1}\) is shaded in green. The position of the circuit at 81 min is plotted in each panel for reference. Letters around each circuit indicate the position of reference parcels specified in Fig. 4.2. All axes are labeled in meters.

The minute period shown. This does not imply that the parcels do not have downdraft origins. At times prior to those shown, some parcels within the circuit are as high as 1500 m AGL. What may be inferred, however, is that parcels reach the surface more slowly than in the control case. This is supported by the finding in Section 3.3 that the low-level downdraft is generally weaker in stable-layer cases.

The 1km10c circuit (Fig. 4.4c) displays altogether different characteristics. Nearly all of the parcels descend from aloft in this case. There also is very little
horizontal convergence of the circuit before 81 minutes. This differing evolution can be explained by a more thorough examination of Fig. 4.2. In the control and 500m5c cases (Figs. 4.2a,b) the vorticity maximum is largely associated with updraft. Thus, in these cases, at least part of the circuit is bound to have origins from a level below 75 m AGL. However, in the 1km10c case (Fig. 4.2c) the low-level vorticity maximum and circuit are mainly embedded in an area of downdraft. Hence, it comes as no surprise that most of this circuit has origins aloft.

It is evident from Fig. 4.3 that a time at which the generation of circulation differs in the control from both the stable layer simulations is approximately 76 minutes. At this time, both the 500m5c case and 1km10c circuit are gradually losing circulation while the control circuit is gaining circulation. This time also is important because the trends in circulation around each circuit here will directly impact the final values of circulation around the locations of the vorticity maxima. Although the physical evolution of each circuit is enlightening in regards to the origin of air parcels that provide low-level circulation to the low-level vorticity maximum, one must return to theory to understand how these differences in circulation generation about each circuit arise. Equation 4.2 implies that circulation is generated or lost where the fluid circuit changes height in areas of buoyancy perturbation. In Fig. 4.5, negative buoyancy integrated over the lowest kilometer is plotted with each circuit at 76 minutes.

The dynamics behind the positive circulation generation in the control at 76 minutes are made clear by Figs. 4.5a,b. Integrating counterclockwise around the circuit, the circuit ascends approximately between parcels F and E (Fig. 4.4a; 76 min) in an area of negative buoyancy (Fig. 4.5b). By Equation 4.2 this requires that negative circulation be generated along this branch of the circuit (in Equation 4.2 \( B \) is negative while \( dz \) is positive, so the change in circulation is negative). However, between parcels E and C the circuit descends back to near the surface in an area of more strongly negative buoyancy. This results in positive circulation generation along that branch of the circuit (in Equation 4.2 both \( B \) and \( dz \) are negative, so the change in circulation is positive). In the rest of the circuit, there is relatively little change in height or no buoyancy perturbation, thus there is no significant contribution to circulation there. Because the change in height is the same over both the ascending and descending branches but the buoyancy is more strongly
negative in the descending branch, the net circulation generation must be positive.

By a similar argument, it can be shown why circulation decreases around the circuit in the 500m5c and 1km10c cases at 76 minutes. In Figs. 4.5c,d and Fig. 4.4b (76 min), the 500m5c circuit exhibits ascending (parcels F to D) and descending (parcels D to C) branches similar to the control case. As in the control case, the ascending branch is in an area of negative buoyancy and leads to negative circulation generation. However, the descending branch occurs in an area of less negative buoyancy than the ascending branch, so the positive circulation generation in that branch is slightly weaker than the negative circulation generation in the ascending branch. Thus, the net change in circulation at this time is slightly negative in the 500m5c case. The 1km10c circuit (Figs. 4.5e,f ; Fig. 4.4c 76 min) also has a weak decrease in circulation at this time because its ascending branch (parcels B to F) generally lies in an area with somewhat more negative buoyancy than its descending branch (parcels F to C).

From Fig. 4.5 one also can see how circulation around the fluid circuits is related to vorticity. Equation 4.3 mandates that the average vorticity normal to the surface bound by the circuit is simply the circulation divided by the area of the surface. As such, one might expect a circuit with positive circulation and vertical excursions to intersect horizontal vorticity vectors. In the control case (Fig. 4.5b), horizontal vorticity vectors oriented toward the south pass through the circuit beneath segment FED in a manner that is consistent with the positive circulation around the circuit. Expectedly, the 500m5c circuit also is intersected by horizontal vorticity vectors under segment FED. As both circuits complete their descent and the surfaces bounded by them become purely horizontal, it is not hard to imagine how these vorticity vectors must be reoriented vertically to satisfy Equation 4.3. In the 1km10c case (Fig. 4.5c), the relative lack of intersection with horizontal vorticity is consistent with the weak circulation that is observed.

Circulation is just a means of tracking the vorticity of many parcels within the broad area bounded by the circuit as it approaches the low-level vertical vorticity maximum. This vorticity may be associated with the ambient vertical wind shear (hereafter referred to as “barotropic” vorticity) or due to solenoidal effects associated with the storm’s internally generated horizontal buoyancy gradients (“baroclinic” vorticity). If the circuits are traced back far enough in time, it stands to
reason that they would have circulation that is purely the result of barotropic vorticity. Equation 4.2 describes how this initial circulation changes as a result of baroclinic generation of vorticity as the circuits encounter the storm. In these simulations, baroclinic vorticity (and circulation) generation plays an essential role in the development of the low-level vertical vorticity maxima because the orientation of barotropic (ambient, horizontal) vorticity intersecting the circuits at a very early time in their evolution actually contributed to negative circulation about the circuits that later encircled the low-level vertical vorticity maxima as they intensified (positive circulation is required for the circuit to enclose positive vertical vorticity at 81 minutes).

It has been shown how positive circulation is acquired by each circuit before 81 minutes and the reasoning behind the sign of circulation generation in each case has also been explained. However, this does not fully explain why the circulation is generally weaker in the simulations with stable boundary layers. The insight provided by this singular analysis suggests that though the circuits may have similar geometries, differences in the spatial distribution of buoyancy perturbations lead to varying circulation generation. At low levels, there are less intense negative buoyancy perturbations in the stable layer simulations because the environment at those levels is already cooled. Though the 500m5c circuit is capable of generating circulation of a magnitude similar to the control, the increased buoyancy gradients at low levels in the control allow circulation to fluctuate more rapidly than in the stable layer simulations. When buoyancy gradients are strong along a material circuit, circulation can be generated or destroyed quickly because the ascending branch of a circuit may be located in an area of significantly less negative buoyancy than the descending branch. Without buoyancy gradients, this is not possible and net changes in circulation around the circuit are limited.

This circulation analysis is likely a reflection of the physical process whereby the presence of a stable layer weakens downdraft intensity at the surface. As discussed in Section 4.1.2, downdraft strength decreases at low levels as a result of decreased negative buoyancy. If the downdraft is weaker, it is likely that weaker horizontal buoyancy gradients exist, thereby decreasing the baroclinic generation of vorticity that has been shown to be necessary to establish the low-level vertical vorticity maximum. Thus, vertical vorticity is likely weaker at low-levels in elevated
supercells because the stable layer decreases the negative buoyancy of downdraft parcels which then hinders the production of baroclinic vorticity that is available to be tilted (the key, of course, is that much of this tilting occurs during the descent of air parcels—a process that also is diminished with a weaker downdraft) and stretched by an updraft.

4.2.3 Convergence of Low-Level Circulation

Up to this point, the discussion has been mainly concerned with what happens before 81 minutes in simulation time. The differences in elevated supercells that lead to decreased low-level circulations have been demonstrated. However, the vorticity maxima at 85 minutes in the control and 500m5c case are the result of convergence of the low-level circulation at 81 minutes.

The positions of the circuits initially surrounding the 81 minute low-level vorticity maxima at 85 minutes are shown in Fig. 4.6. It is clear that after 81 minutes the control (Fig. 4.6b) and 500m5c (Fig. 4.6d) circuits have experienced a convergence of their horizontal area. By Equation 4.3 this requires an increase in the vertical vorticity bounded by the circuit because circulation remains relatively constant as the limited horizontal span of the circuit means that its vertical excursions occur within similar values of buoyancy. The horizontal area of the circuit at 85 minutes in the control is smaller than in the 500m5c case; thus, it can be inferred that vorticity increases less within the 500m5c case. Physically, this difference occurs because the low-level updraft is weaker in the stable boundary layer case. Through continuity, a weaker updraft implies less convergence at the surface, resulting in less convergence of the circuit. Figs. 4.6a,c also show less vertical motion of the circuit within the 500m5c case. This is a result of the weaker updraft, and suggests that less stretching of vorticity is possible in the stable boundary layer case.

The 1km10c circuit (Figs. 4.6e,f) has a noticeably different behavior than its less stable counterparts. Instead of converging within an updraft, this circuit is forced towards the ground by downdraft and diverges. Thus, if circulation remains relatively constant (which it must since there is a limited vertical component to the circuit) the average vertical vorticity within this circuit decreases. In terms of vorticity, the stretching term in this situation destroys vertical vorticity rather
than enhances it.
Figure 4.5.  (a,c,e) Vertical cross-sections of the material circuits (red) at 76 minutes looking from the south.  (b,d,f) Plan view of circuits (red) at 76 minutes (with reference letters) and 81 minutes. Contours of negative buoyancy integrated over the lowest 1 km are plotted and shaded according to the colorbar with units of m s$^{-2}$. Arrows are horizontal vorticity vectors at 250 m AGL (indicated in a,c,e by the black line) with arrow length scaled by magnitude.
Figure 4.6. The forward evolution of material circuits is displayed. (a,c,e) Similar to Fig. 4.5a,c,e but with the circuits shown at 85 minutes. (b,d,f) Similar to Fig. 4.2a,b,c but without vertical vorticity contours and the circuits shown at 81 minutes (circular) and 85 minutes (irregular shape).
Chapter 5

Summary and Conclusions

Motivated by the prevailing notion that elevated supercells are less likely to produce tornadoes than surface-based supercells, a three-dimensional numerical model was used to determine the effects of a statically stable boundary layer on airflow and low-level rotation in supercell thunderstorms. The results of a series of simulations, each with a different degree of low-level static stability, were compared against those of the control experiment lacking a significantly stable boundary layer. Storms were initiated with a warm bubble in an environmental temperature profile in which the lowest layer was modified by a linear reduction in potential temperature between the top of the stable layer and the ground.

As the depth and surface temperature deficit of each stable layer are adjusted, surface-based CAPE is decreased by surface cooling, while most unstable CAPE remains constant. Over the parameter space of simulations, as the CIN increases in each simulation, it was found that near-ground vertical vorticity and magnitudes of vertical velocity (both updraft and downdraft) decrease. At a higher level (e.g., 4 km AGL) within the storm, however, the imposition of even the strongest stable layer has little effect on these quantities. Near-surface vertical vorticity and downdraft magnitude were also found to decrease as DCAPE decreases, suggesting a link between downdraft intensity and near-surface rotation. It also was found that the strength of near-surface rotation was positively correlated with low-level updraft magnitude.

The results also suggest that the existence of a stable boundary layer does not necessarily require that a supercell located above it be elevated. It was found that
in all cases but those with the most stable surface layer, storms are quite capable of drawing inflow from near the surface. Even in those cases with the deepest and coldest boundary layers, some updraft parcels still originated well within the stable boundary layer. As such, the classification of storms as strictly elevated or surface-based is deemed inappropriate, and a continuum approach suggested by Corfidi et al. (2008) is supported. Downdraft airflow was also found to be less affected by the presence of a stable boundary layer than previously surmised. In all cases, downdraft parcels originating at 2 km AGL are able to reach the surface despite reduced negative buoyancy or even a period of positive buoyancy along their descent. However, the magnitude of vertical velocity within downdraft regions generally decreases at low levels with increasing stable boundary layer strength because DCAPE decreases.

As a means of explicating the dynamical processes most responsible for decreased low-level rotation in supercells in environments with surface stability, the circulation around material circuits encircling the low-level vorticity maxima was computed. It was found that the circulation is generally weaker around the low-level vorticity maxima as boundary layer stability increases. When circuits were followed backward in time from a particular low-level vorticity maximum, it was found that the control case was generally able to generate more circulation because of greater negative buoyancy within the storm cold pool. In contrast, the purely elevated case has very weak negative buoyancy and differing circuit geometry such that generation of positive circulation is reduced. An intermediate case in which the storm is not truly elevated also was analyzed. In that simulation, it was found that differences in circulation generation from the control are less easily generalized than in the truly elevated case.

The convergence of low-level vorticity by the storm updraft also was investigated. In the control case, it was found that low-level convergence associated with the updraft is greater, and thus vertical vorticity is more amplified than in an intermediate case. In an environment with a strong temperature inversion near the surface in which the storm is fully elevated, vorticity was unable to be converged and stretched by an updraft because the investigated vertical vorticity maximum occurred principally in downdraft.

Primary findings and conclusions drawn from these results as they relate to
real supercells are summarized as follows:

1. Storms located above weak to moderately stable boundary layers may not be truly elevated, in the sense that these storms still draw inflow from the statically stable layer in contact with the surface.

2. The existence of even a strong stable boundary layer does not prevent downdrafts from reaching the surface, but downdraft speed is decreased through reduced negative buoyancy.

3. Supercells situated over a stable boundary layer such as those found at night or on the cool side of a front have weaker low-level vertical velocity and weaker near-ground rotation.

4. Low-level vertical vorticity is weaker in elevated supercells because of both decreased generation of baroclinic vorticity and decreased convergence and stretching of low-level vertical vorticity.

5. Baroclinic vorticity generation is weaker in elevated supercells because downdrafts are less negatively buoyant with respect to the low-level environment.

6. Low-level convergence of vertical vorticity is weaker because increased CIN and decreased surface-based CAPE in elevated supercells decreases the strength of low-level updrafts.

7. Differences in vertical vorticity and vertical velocity within the simulated elevated supercells are limited to low levels. At midlevels, the elevated supercells display similar traits to the surface-based storms.

In light of these conclusions, it can be said with a fair amount of confidence that elevated supercells are indeed less likely to produce tornadoes. As previously mentioned, the grid spacing of the simulations in this study is too large to resolve a tornado. However, it has been shown that the near-ground circulation that precedes tornadogenesis is significantly reduced in truly elevated supercells. Supercells that are located over stable boundary layers yet still remain surface-based also show a tendency toward decreased near-ground circulation, thus it is probable that the development of a tornado may be less likely in these situations as well.
Some recent field observations suggest that baroclinic vorticity may not be as important to tornadogenesis as the results of modeling studies imply. Shabbott and Markowski (2006) found reduced baroclinity in tornadic supercells when compared with nontornadic supercells. Others have found that observed temperatures within the rear-flank downdraft are generally warmer in tornadic supercells than in nontornadic storms (Markowski et al. 2002; Grzych et al. 2007). As previously discussed, the thermodynamic characteristics of downdrafts and cold pools in simulations show sensitivity to the microphysical parameterization that is used (cf. Section 2.1). Thus, it is possible that the importance of baroclinic vorticity in the development of the low-level circulations may be exaggerated by the model in this study. Baroclinic vorticity generation, although perhaps not as dominant in real storms as in modeled storms, still seems to be significant in real storms per the vortex line analyses of Markowski et al. (2008). Regardless of whether the most important baroclinic generation is in the forward- or rear-flank downdraft region, it seems likely that a stable layer at low levels would have a suppressing effect on baroclinic generation within descending parcels.

It is likely that more variables than simply the thermodynamic characteristics of a homogeneous stable layer influence the development of low-level rotation in elevated supercells. Wind shear and humidity (both relative and absolute) modifications that often accompany the development of statically stable low levels likely also are important. The scenario modeled herein most closely resembles the situation of a supercell developing over a nocturnal boundary layer. This situation was chosen for its simplicity and the ability to focus purely on the effects of altering the low-level thermodynamic profile. In the future, a more realistic configuration might simulate the interaction of a preexisting storm with a developing nocturnal boundary layer. Furthermore, these simplified simulations are conducted in a horizontally homogenous environment, whereas it has been found that storm interactions with (horizontally inhomogeneous) mesoscale boundaries often increase the tornadic potential of supercells. As such, subsequent simulations might investigate the case in which a storm crosses or propagates along a boundary between an unstable air mass and the stably-stratified outflow of a previous storm.

Most of the observed differences between the simulated surface-based and elevated supercells occur in close proximity to the ground. As such, key differences in
storm structure that might reveal the tornadic potential of a thunderstorm would typically be well below the scanning horizon of a WSR-88D (Weather Surveillance Radar 1988 Doppler). Therefore, nowcasting tornado potential will remain difficult in the short term even with an increased understanding of how near-ground rotation is inhibited in elevated supercells. Nowcasting improvements likely await a better understanding of how much low-level CIN a supercell can encounter before the likelihood of tornadogenesis is significantly diminished, how such thresholds vary as a function of the supercell environment (e.g., vertical wind and buoyancy profiles), and better real-time observations of storm environments.
References


