Vertical Variations of Boundary Layer Potential Buoyancy in Tornadic and Nontornadic Near-Storm Environments

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ABSTRACT

Despite great strides in understanding the tornadic near-storm environment (NSE), at times it remains difficult to determine why some storms produce significant tornadoes, while others produce none, given similar pretornadic radar reflectivity and velocity signatures. Previous studies have shown that this is likely related to the potential buoyancy ($u_{ep}$) of the rear-flank downdraft (RFD) air. Unfortunately, to date there are few ways to operationally anticipate possible RFD thermodynamic character. Based upon previous research indicating that capping inversions may restrict much of the low-level RFD air to come from within the boundary layer, this study considers the relation of $\Delta u_{ep}$ (vertical change in $u_{ep}$ within the boundary layer below the cap) to tornadogenesis potential. This is because when a cap exists above a boundary layer and the descent of lower-$\theta_e$ air from aloft to the surface is potentially limited, then minimal $\Delta u_{ep}$ may indicate more RFD air that has greater potential buoyancy. The Rapid Update Cycle (RUC) soundings used in this study and several observed soundings taken in the vicinity of violent tornadoes suggest that boundary layer $\Delta u_{ep}$ shows promise as an additional means of discriminating between tornadic and nontornadic NSEs.

1. Introduction

Recent studies of tornado-producing environments have indicated that strong low-level wind shear and high values of near-surface moisture are critically important to tornadogenesis (e.g., Davies-Jones et al. 1990; Johns et al. 1993; Rasmussen and Blanchard 1998, hereafter RB98). This is reflected in the study of parameters like 0–1-km bulk shear and lifted condensation level (LCL) height, both of which have shown promise in discriminating between tornadic and nontornadic near-storm environments (NSEs) in studies such as that by Thompson et al. (2003, hereafter T03). The importance of boundary layer characteristics may be related to the role of the rear-flank downdraft (RFD; Lemon and Doswell 1979) in tornadogenesis.

Markowski et al. (2002, hereafter M02) found that surface RFD air in cases of significant (F2 and greater intensity) supercellular tornadoes tended to have greater potential buoyancy compared to surface RFD air associated with nontornadic supercells. Specifically, M02 found that surface RFD equivalent potential temperature ($\theta_e$) deficits relative to the environment were typically less than 4 K in significant tornado cases, and greater than 10 K in nontornadic events in their study.

Askelson (2002) and Askelson et al. (2004, hereafter A04) built upon this work by investigating the influence of thermodynamic structures on simulated low-level downdraft $\delta_e$ values using a 1.5-dimension, hydrometeor-driven downdraft model. Note that in both M02 and A04 (as well as the present study), $\theta_e$ is actually the pseudoequivalent potential temperature ($\theta_{ep}$; Emanuel

1 Bulk shear refers to the magnitude of the vector wind difference.

2 The enhanced Fujita (EF) scale has been used in place of the Fujita (F) scale in the United States since 2007.
1994, hereafter E94) computed using the formula derived by Bolton (1980). A04 performed simulations using various thermodynamic structures and concluded that a capping inversion (i.e., a statically stable layer with temperature lapse rates less than moist adiabatic at the top of the boundary layer) can prevent low-$\theta_{ep}$ air from above the cap from descending to the surface. This can in turn minimize surface $\theta_{ep}$ deficits (depending on boundary layer $\theta_{ep}$ structure), which, based on the M02 study, is favorable for tornadogenesis. Given the presence of a cap in the prescribed environments of the A04 simulations, simulated downdrafts were either generated or intensified significantly in the boundary layer as a result of the dry-adiabatic temperature lapse rate typically present beneath the cap. This, too, is consistent with the observations of M02, who found that surface RFD air often originated at an altitude $\leq$ 1 km above ground level (AGL), especially in tornadic cases. Expanding this line of research further, Naylor et al. (2012) used a fully compressible three-dimensional model [the Weather Research and Forecasting Model (WRF)] to investigate the influence of a cap on downdrafts and found that in their capped simulation surface cold-pool air parcels have larger $\theta_{ep}$ values and lower origination heights than surface cold-pool air parcels in their non-capped simulation.

Given the results of A04 and Naylor et al. (2012), it is not immediately obvious as to why caps can be effective at inhibiting air from above the boundary layer from reaching the surface. If one examines downdraft convective available potential energy (DCAPE) for parcels residing above a capped boundary layer, for instance, significant DCAPE is commonly found to exist for many of those parcels. DCAPE, however, is an idealization of the amount of energy gained by actual downdrafts and, as such, represents, at best, an indication of downdraft potential (Gilmore and Wicker 1998, hereafter GW98).

Downdrafts generally realize only a fraction of their potential acceleration (E94, p. 172; GW98). This occurs because of two primary reasons: 1) the rate of compressional heating generally exceeds that of cooling owing to hydrometeor evaporation and melting and, thus, downdrafts descend between dry and moist adiabatically [as opposed to the moist-adiabatic descent assumed with DCAPE; Das and Subba Rao (1972); E94] and 2) downdrafts immediately start to sink as soon as they become negatively buoyant and, thus, their starting temperatures are not their isobaric wet-bulb temperatures, as is assumed with DCAPE (E94; GW98).

GW98 provide an excellent illustration of both of these effects in their Fig. 7 (not shown), in which downdrafts, throughout most of their descent, realize very small downward accelerations owing to temperature differences. Near the surface, where parcels may travel significant distances horizontally while experiencing hydrometeor-driven cooling, downward acceleration owing to temperature differences was enhanced (GW98). While additional effects including enhancement owing to hydrometeor loading, weakening owing to entrainment, and perturbation pressure forcing are important to overall downdraft strength, the numerical results of GW98 (their Fig. 3), which include these processes, show that downdrafts commonly effectively realize only a small fraction of their DCAPE ($\approx$ 30%). Thus, downdrafts, especially in environments that are not dry adiabatic [which is the most permissive of downdrafts; Srivastava (1985, 1987)], commonly descend with relatively small amounts of negative buoyancy. Caps, in which the negative buoyancy of downdrafts is significantly reduced owing to the inherent change in lapse rate present within them, consequently, can be expected to have significant impacts on downdrafts, including possibly halting them altogether.

Because of the simulated impacts of caps on downdrafts, A04 concluded that the boundary layer $\theta_{ep}$ profile may be important to surface RFD $\theta_{ep}$ deficits (and thus tornadogenesis), and suggested that the vertical lapse rate of boundary layer $\theta_{ep}$ should be investigated further. The purpose of the present study is to determine whether differences in low-level $\theta_{ep}$ profiles exist between significantly tornadic and nontornadic environments that would be suitable for use by operational forecasters. Details regarding the methodology and data used in this study are presented in section 2, and results are provided in section 3. The results are examined along with potential failure modes in section 4, and our findings are summarized in section 5.

2. Methodology and data

The model simulations performed by A04 indicated that when environmental $\theta_{ep}$ does not decrease with height in a boundary layer that is capped, lower-$\theta_{ep}$ air is much less likely to descend to the surface. Consistent with A04, in the present study the boundary layer is defined as the near-surface layer with unstable or conditionally unstable temperature lapse rates, while the capping inversion is defined, based on the American Meteorological Society’s (AMS) Glossary of Meteorology (Glickman 2000), as the first statically stable zone above the boundary layer (i.e., the first layer with temperature lapse rates less than moist adiabatic). In contrast, when $\theta_{ep}$ decreases significantly in the boundary layer, where the downdraft tends to develop and accelerate, those simulations indicated that lower $\theta_{ep}$ values from within the boundary layer could readily descend to
the surface. The focus of this study is therefore on the change of \( \theta_{ep} \) from the surface to the base of the cap, referred to as boundary layer \( \Delta \theta_{ep} \).

Proximity soundings are commonly used in the study of tornado environments (e.g., Showalter and Fulks 1943; Fawbush and Miller 1954; Darkow 1969; RB98). This approach has proven to be invaluable in helping to understand thermodynamic and kinematic structures in the vicinity of tornadoes. However, many challenges complicate the task of collecting representative observed soundings, most notable of which is obtaining a sufficient number of observed soundings that are representative or in “close proximity” (T03). As a result, recent studies such as T03 have focused on the use of Rapid Update Cycle (RUC) gridpoint analysis (0-h forecast) soundings as a surrogate in the severe convective environment. Since these soundings are not tied to an existing upper-air station network and synoptic hour or occasional special observations, they enable studying a much greater number of events, even though the RUC analysis soundings tend to be a little too cool and dry at the surface (T03). This study also uses RUC analysis thermodynamic soundings extracted from analyses having 40-km horizontal grid spacing, as described by Benjamin et al. (2002). Kinematic data were not available from these soundings, so plan views of RUC analysis data at 40-km grid spacing from the Storm Prediction Center (SPC) analysis scheme (Bothwell et al. 2002) were used to derive rough estimates of shear-related parameters. The shear-related parameters, obtained from horizontal contour maps, are only known to within 2.6 m s\(^{-1}\) (i.e., 5 kt) for the 0–1-km bulk shear and 50 m\(^2\) s\(^{-2}\) for the 0–1-km storm-relative helicity (SRH) due to the contour intervals used to generate the plots. SRH values rely on the internal dynamics method of supercell motion developed by Bunkers et al. (2000).

Values for significant tornado events are compared to those for cases involving tornadic false alarms, wherein tornado warnings were issued but no tornado occurred. This delves into the heart of the warning decision-making problem by directly seeking out storms that were significant enough from a radar and environmental standpoint to garner public warnings but which did not produce tornadoes. To avoid isolated instances, nontornadic events are characterized by at least two unverified tornado warnings within a 6-h period across a National Weather Service (NWS) county warning area (CWA). To develop a representative sample of nontornadic events, cases that produce even a single tornado in the CWA, of any magnitude, are not used. Note that CWAs are used in this analysis to prevent a spatially large convective outbreak containing both nontornadic and significantly tornadic cases from being completely removed from the event database.

Tornadic false alarm cases from January through December 2008 across the central United States are used in this study, as gathered from the internal NWS verification database (NWS Performance Management web site; https://verification.nws.noaa.gov). Cases were retrieved from the CWAs shown in Fig. 1. Some small bias may exist in the nontornadic sample given the possibility that in some cases a tornado could have occurred without being reported. A database of tornadic cases was created by collecting data from the National Climatic Data Center’s Storm Data publication regarding significant (EF2–EF5) tornadoes from the same time period and areas as the nontornadic cases. Only EF2–EF5 tornadoes are considered in the database in order to further discriminate between cases with failed tornadogenesis and those that produce the most dangerous tornadoes.

Once nontornadic and significantly tornadic cases were identified, subjective radar analysis was performed to classify each event as either an isolated, discrete supercell (DS), or a multiple-cell (MC) event. The MC category is defined as convection without any significant breaks in the precipitation field between cells (i.e., elements of the convective system are connected by reflectivity of at least around 30 dBZ). The MC category therefore includes linear modes of convection such as squall lines, but may also contain embedded supercell structures. Since convective mode is a known contributor to tornadogenesis failure (Thompson and Mead 2006), this subjective categorization was used to separately assess statistics for isolated, strongly rotating DSs that ordinarily portend a higher tornado threat than their MC counterparts. Given the propensity for convective episodes to contain an ever-changing spectrum of both types, events were classified based on whichever
type resulted in the highest-rated tornado or most tornado warnings.

The process of visually inspecting each storm assured that each of the RUC gridpoint soundings used in the study was taken in the “inflow” region of those storms. Quality assurance was performed to ensure that the soundings are not convectively contaminated. Any soundings not containing any CAPE were removed. The soundings for the nontornadic events were retrieved from the storm’s inflow region for the analysis hour and grid point closest to the subjectively determined peak intensity of the storm, which was based on traits such as the reflectivity maximum, the inflow-side reflectivity gradient along the hook echo, the presence of an inflow notch and/or gust front wrapping into the notch, and the location of the reflectivity maximum relative to the inflow notch and hook echo. For the significantly tornadic cases, soundings were taken from the inflow side of storms and as close as possible in time and space to the highest-rated tornado of an event. Although the process of individually inspecting data for each event reduced the total number of cases used in the study, it did rule out the possibility of soundings being taken in the poststorm environment in which CAPE can be completely removed by deep convection or a boundary passage. Since the focus of this study is on the NSE of each individual event, and all of the nontornadic cases had tornado warnings during the time of sounding analysis, only subjective velocity interpretation was used in determining peak intensity. It is acknowledged that this may result in some bias to the results.

The following thermodynamic and kinematic parameters were computed for each nontornadic and significantly tornadic case: 100-hPa-mean parcel CAPE (MLCAPE), 100-hPa-mean parcel LCL height (MLLCL), 100-hPa-mean parcel level of free convection (MLLFC), 0–1-km bulk shear, and 0–1-km SRH. CAPE values are calculated using the virtual temperature correction described by Doswell and Rasmussen (1994). These parameters were computed in order to determine the representativeness of this relatively small sample of cases based upon statistics gathered from larger RUC-based studies (such as T03). Finally, boundary layer $\Delta \theta_{ep}$ values were computed for each sounding by subtracting the value of $\theta_{ep}$ at the surface from the value of $\theta_{ep}$ at the upper bound of the boundary layer, which corresponds to the lower boundary of the cap. Example calculations of boundary layer $\Delta \theta_{ep}$ are shown in Figs. 2 and 3.

![Fig. 2. Skew $T$–log$p$ diagram for the 0000 UTC 18 Jul 2011 Bismarck, ND, observed sounding showing an example of how to calculate boundary layer $\Delta \theta_{ep}$. Solid lines indicate environmental temperature profiles and dashed lines indicate environmental dewpoint temperature profiles. Temperatures and dewpoint temperatures are in °C and pressure is in hPa.](image-url)
Note that as long as a definitive upper bound of a boundary layer could be defined, such a sounding was used even if the temperature lapse rate within the boundary layer was significantly less than dry adiabatic. This is because 1) an RFD can still exist when lapse rates are not dry adiabatic (Das and Subba Rao 1972; Srivastava 1985, 1987) and 2) observed soundings taken in close proximity to violent (EF4 and EF5) tornadoes were often characterized by less than dry-adiabatic lapse rates that existed beneath a distinct cap. These observations also support analyzing the change in boundary layer $\theta_{ep}$, rather than just the water vapor mixing ratio lapse rate, which would be appropriate only when a dry-adiabatic temperature lapse rate is present throughout the whole boundary layer.

A deep stable layer at the surface is not conducive to thermodynamically driven formation, intensification, or maintenance of downdrafts, which is a primary basis for the hypothesis being tested here. Therefore, boundary layer $\Delta \theta_{ep}$ values were calculated separately for soundings that contained a surface-based stable layer, such as the one in Fig. 4. In these special cases, the surface-based
stable layer was considered to be part of the boundary layer for the practicality of calculating \( \Delta \theta_{ep} \), and the top of the boundary layer was defined as the first absolutely stable layer above the surface-based stable layer.

Once cases with missing sounding data or zero-CAPE soundings were removed from the dataset, 33 non-tornadic and 21 significant (EF2–EF5) tornado cases without surface-based stable layers were available for analysis. Based on subjective radar analysis of cases without surface-based stable layers, 22 (≈67%) of the nontornadic and 15 (≈71%) of the significant tornado cases were DSs. An additional 9 nontornadic and 10 significantly tornadic RUC soundings contained surface-based stable layers. Results for boundary layer \( \Delta \theta_{ep} \) were calculated separately for the events having surface-based stable layers (including surface-based inversions). Three (≈33%) of the nontornadic and 0 (0%) of the significant tornado cases with surface-based stable layers were DSs. Table 1 summarizes the cases included in the study and their categorization based on storm type and whether or not they included a surface-based stable layer. These cases are hereafter referred to as the T07 dataset.

### Table 1. A summary of the number and categorization of cases used in the study.

<table>
<thead>
<tr>
<th>Category</th>
<th>No. of nontornadic cases</th>
<th>No. of significantly tornadic (EF2–EF5) cases</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>All cases without surface-based stable layers</td>
<td>33</td>
<td>21</td>
<td>54</td>
</tr>
<tr>
<td>Multiple-cell (MC) cases</td>
<td>11</td>
<td>6</td>
<td>17</td>
</tr>
<tr>
<td>Discrete supercell (DS)</td>
<td>22</td>
<td>15</td>
<td>37</td>
</tr>
<tr>
<td>All cases with surface-based stable layers</td>
<td>9</td>
<td>10</td>
<td>19</td>
</tr>
<tr>
<td>MC cases</td>
<td>6</td>
<td>10</td>
<td>16</td>
</tr>
<tr>
<td>DS cases</td>
<td>3</td>
<td>0</td>
<td>3</td>
</tr>
</tbody>
</table>
Finally, a small group of observed soundings taken within approximately 115 km and 75 min of violent tornado events were used to provide some validation of the use of boundary layer $\Delta \theta_{ep}$ from an observational standpoint. The observed soundings were extracted from the inflow region of the violent-tornado-producing storms. The spatial and temporal requirements of being within 115 km and 75 min of violent tornado events are similar to those employed by Davies and Johns (1993) and Kerr and Darkow (1996).

3. Results and analysis

a. Results for commonly used thermodynamic and kinematic parameters

For this dataset there is significant spread and overlap in MLCAPE between nontornadic and significant tornado cases (Fig. 5a). The operationally insignificant discrimination across categories shown here is consistent with the results of RB98 and T03.

The height of the MLLCL showed separation between nontornadic and significant tornado events (Fig. 5b), with an MLLCL height of $\sim 900$ m AGL discriminating between categories. RB98 and T03 also found that LCL height is a discriminator between nontornadic and significant tornado cases. MLLCL results in the present study are similar to T03 for their nontornadic supercells, but the MLLCL heights for the significant tornado cases are slightly lower than in T03. The box-and-whiskers diagrams showing MLLFC height revealed substantial overlap, thereby suggesting that it is not very useful for discriminating between nontornadic and significant tornado events (Fig. 5c).

Shear-related parameters also strongly discriminated between the two categories of events considered in this database, which is again similar to the results of RB98 and T03. The 0–1-km bulk shear and 0–1-km SRH showed no overlap between the 25th and 75th percentiles of nontornadic and significantly tornadic cases (Figs. 5d and 5e). In fact, only the minimum values of 0–1-km bulk shear and 0–1-km SRH associated with significant tornado cases were near the 60th percentile of the nontornadic events.

Although the exact numerical values and specific degree of parameter overlap for nontornadic and significantly tornadic cases in this study differ somewhat from larger data samples such as those considered by RB98 and T03, they show similar trends. This increases confidence that the RUC sample herein is indeed representative even if it is smaller than those used in other studies as a result of the need to investigate each case individually.

b. Results for boundary layer $\Delta \theta_{ep}$

Boundary layer $\Delta \theta_{ep}$ results for soundings without surface-based stable layers are shown in the
and significantly tornadic DS cases (means of $-6.3$ and $-0.4$ K, respectively) is also statistically significant at the 99% confidence interval. Thus, $\Delta \theta_{ep}$ appears to have utility in discriminating tornado potential, especially in DS environments.

Boundary layer $\Delta \theta_{ep}$ was also calculated for each of the cases for which the associated RUC sounding profile has a surface-based temperature stable layer. The sample size of these cases is small, but significant overlap exists between the 25th and 75th percentiles of these non-tornadic and significant tornado cases (not shown). Analysis of individual cases suggests that even when the mixing ratio decreases with height in the lower part of the profile, the negative lapse rates within the surface-based temperature stable layer often dominate the $\theta_{ep}$ profile such that the boundary layer $\Delta \theta_{ep}$ value is relatively high. However, stable lapse rates could inhibit transport of higher $\theta_{ep}$ above the ground to the surface. Thus, boundary layer $\Delta \theta_{ep}$ may not be particularly useful in these instances, where anomalously strong SRH or bulk shear could overcome the negative influence of nocturnal, surface-based inversions (Davies and Fischer 2009). It is also worth noting that the appearance of surface-based stable layers in some of these cases could merely be an artifact of unrepresentative RUC thermal profiles.

RB98 had success in combining thermodynamic and kinematic parameters in their analysis. Thus, it is worthwhile to compare boundary layer $\Delta \theta_{ep}$ with 0–1-km bulk shear and 0–1-km SRH, since all three appear to strongly discriminate between the two categories of events in this study. Figure 7a shows a comparison of $\Delta \theta_{ep}$ and 0–1-km bulk shear, while Fig. 7b shows a comparison of $\Delta \theta_{ep}$ and 0–1-km SRH for cases without surface-based stable layers. Seventy-one percent of the significant tornado cases in this study had a boundary layer $\Delta \theta_{ep}$ value of $-3$ K or greater and 0–1-km bulk shear of 10 m s$^{-1}$ or greater, and 76% of the significant tornado cases had a boundary layer $\Delta \theta_{ep}$ value of $-3$ K or greater and 0–1-km SRH of 150 m$^2$ s$^{-2}$ or greater.

Finally, $\Delta \theta_{ep}$ values for all the cases were calculated using static layers from 1 to 6 km AGL in order to determine if they could provide any discrimination ability similar to that determined by using a defined boundary layer. This test was also used to investigate the possibility that lower-$\theta_{ep}$ air from above the cap may be descending to the surface in some cases. The results of this analysis are shown in Fig. 8. As seen in the box-and-whiskers plot, the only static layer that demonstrates any discrimination ability is 1 km AGL, where a Student’s $t$ test showed that the difference in means is significant at the 90% confidence interval ($\alpha = 0.10$). The remainder of the layers used to calculate $\Delta \theta_{ep}$ display considerable overlap between significant tornado and non-tornadic

**FIG. 6.** As in Fig. 5, but for boundary layer $\Delta \theta_{ep}$ of (a) 21 significant tornado and 33 nontornadic events and (b) 15 significant tornado and 21 nontornadic events associated only with discrete supercells [a subset of those illustrated in (a)].

Box-and-whiskers diagram in Fig. 6a. Negligible overlap exists between the 25th and 75th percentiles of the non-tornadic and significant tornado cases. It appears that $\Delta \theta_{ep}$ values greater than $-3$ K are favorable for the development of significant tornadoes. In contrast, the nontornadic cases are associated with $\Delta \theta_{ep}$ less than $-3$ K and, more often than not, with values less than $-4$ K. This is consistent with the hypothesis of A04 that minimal change in, or an increase of, $\theta_{ep}$ within the boundary layer leads to smaller $\theta_{ep}$ deficits within RFDs and thus increased odds of tornadogenesis (based on M02). A Student’s $t$ test was also applied to the boundary layer $\Delta \theta_{ep}$ results shown in Fig. 6a using a two-tailed test and assuming unequal variances (Milton and Arnold 1990). This test revealed that the difference in means between nontornadic and significantly tornadic cases (means of $-5.8$ and $-1.4$ K, respectively) is statistically significant at the 99% confidence interval ($\alpha = 0.01$).

To better understand the utility of $\Delta \theta_{ep}$ for different storm modes, box-and-whiskers diagrams were also generated for the subset of MC and DS cases without surface-based stable layers in the database, as outlined in Table 1. Significant overlap exists between the boundary layer $\Delta \theta_{ep}$ values of nontornadic MC cases and significantly tornadic MC cases (not shown). However, boundary layer $\Delta \theta_{ep}$ actually shows even more separation for DS cases than when considering all modes (Fig. 6b). The 75th percentile $\Delta \theta_{ep}$ of the nontornadic cases is $-2.8$ K, while the 25th percentile $\Delta \theta_{ep}$ of the significant tornado cases is $-1.4$ K. The difference in means between nontornadic
cases, and Student’s $t$ tests suggest no significant difference in the means.

4. Discussion

a. Additional observations of boundary layer $\Delta \theta_{ep}$

A sample of RUC analysis soundings used in T07 and taken within 40 km and 30 min of radar-identified supercell events was used to calculate boundary layer $\Delta \theta_{ep}$ for a larger sample of events from across the conterminous United States. Soundings without CAPE or which contained surface-based stable layers were not used in order to assure that the same methodology was applied to these data. Boundary layer $\Delta \theta_{ep}$ was thus calculated for 103 significant tornado cases, 243 weak (F0 or F1 intensity) tornado cases, and 371 nontornadic cases using the T07 dataset. The results of this analysis are shown in Fig. 9.

From Fig. 9 it is apparent that the significant tornado soundings from T07 occupy a parameter space very similar to the one associated with the dataset presented in section 3 and shown in Fig. 6. A Student’s $t$ test revealed that the difference in means between the T07 sample ($-2.4$ K) and the dataset presented in section 3 ($-1.4$ K) is not significant at the 99% confidence interval. When considering only significant tornado cases in the T07 sounding set with $0$–$1$-km bulk shear $\geq 10$ m s$^{-1}$, the mean boundary layer $\Delta \theta_{ep}$ is actually $-0.2$ K. In addition, the 25th–75th percentiles of boundary layer $\Delta \theta_{ep}$ values for the weak tornado cases closely matches those of the significant tornado cases from both sets of data. The mean value of boundary layer $\Delta \theta_{ep}$ for the weak tornadoes from the T07 sounding sample is $-2.9$ K.

The nontornadic cases in Fig. 9 display substantially more overlap with the significant tornado events than that shown by the box-and-whiskers plot in Fig. 6. The ability of $\Delta \theta_{ep}$ to discriminate between significantly tornadic and nontornadic cases is thus called into question. However, based on Student’s $t$ tests that indicate the difference in the means for the nontornadic cases between the two datasets is significant at the 99% confidence interval, the nontornadic cases in Figs. 6 and 9 are unlikely to have been obtained from the same underlying populations. This is likely attributable to the differences in case selection methodology between T07 and the soundings collected for the data presented in section 3. Nonetheless, many of the nontornadic cases in Fig. 9 are still associated with boundary layer $\Delta \theta_{ep}$ less than $-3$ K. The mean value for the nontornadic cases is $-4.9$ K. In addition, for the T07 dataset the difference in means between both the significant and weak tornado events and the nontornadic cases is significant at the 99% confidence interval.

The RUC analysis soundings from both datasets contain a wide range of MLCIN values, which themselves display little discrimination between nontornadic and significant tornado cases (not shown). However, the

![Fig. 7. Scatter diagrams of (a) 0–1-km bulk shear (m s$^{-1}$) and boundary layer $\Delta \theta_{ep}$ (K) and (b) 0–1-km SRH (m$^2$ s$^{-2}$) and boundary layer $\Delta \theta_{ep}$ (K) from RUC analysis soundings for cases without surface-based stable layers. Shaded squares represent nontornadic cases and open circles represent significant tornado events.](image-url)
distribution of MLCIN between cases with boundary layer $\Delta\theta_{ep}$ less than $-3$ K and cases with $\Delta\theta_{ep}$ greater than $-3$ K shown in Fig. 10 suggests that cases with minimal $\Delta\theta_{ep}$ deficits in the boundary layer are at least slightly more likely to have MLCIN $\geq 25$ J kg$^{-1}$. Soundings without substantial MLCIN may be more likely to contain boundary layers that extend to several kilometers AGL, where lower-$\theta_{ep}$ air may reside. However, calculated correlation coefficients for all cases ($r = 0.04$ for the data in section 3 and $r = 0.15$ for the T07 data) revealed that the correlation between $\Delta\theta_{ep}$ and MLCIN is not significant at the 95% confidence interval ($\alpha = 0.05$).

To explore the observational characteristics of boundary layer $\theta_{ep}$ profiles, several observed soundings not used in the other analyses of this study and taken in close temporal and spatial proximity to violent tornadoes were collected. Boundary layer $\Delta\theta_{ep}$ was calculated for each of these soundings (Table 2). Most of the observed violent tornado soundings have rich boundary layer moisture, a dry- or nearly dry-adiabatic temperature lapse rate beneath a distinct capping inversion, and water vapor mixing ratios that often increased or remained constant with height in the boundary layer (not shown). The mean boundary layer relative humidity for the sounding set was 87% and the mean boundary layer temperature lapse rate was $7.3^\circ$C km$^{-1}$. The $\Delta\theta_{ep}$ values from the observed soundings also strongly support the hypothesis that boundary layer $\Delta\theta_{ep}$ is greater than $-3$ K in the vicinity of most significant tornadoes (only one sounding had a value less than $-3$ K). In fact, the mean boundary layer $\Delta\theta_{ep}$ of this sample is actually positive at 1.0 K.

Only the 0000 UTC 9 May 2003 Norman, Oklahoma (KOUN), sounding had an outlier boundary layer $\Delta\theta_{ep}$ value of $-5.1$ K (Fig. 11). In fact, if that sounding is removed from Table 1, then the mean boundary layer $\Delta\theta_{ep}$ for these violent tornado events is actually 1.8 K. In
the 0000 UTC 9 May 2003 KOUN profile, even though moisture was rich throughout the boundary layer, the dewpoint decreased from 23°C at the surface to 18°C at the base of the cap near 880 hPa (with a resultant drop in \( u_{ep} \) from 364.2 to 359.1 K). Note, however, that otherwise this sounding indicates favorable conditions for violent tornado development, with 0–1-km bulk shear of \( \sim13 \) m s\(^{-1}\), 0–1-km SRH over 300 m\(^2\) s\(^{-2}\) based on observed storm motion, and an MLLCL height of \( \sim1150 \) m AGL. This sounding also indicates very large instability, with MLCAPE of \( \sim5000 \) J kg\(^{-1}\). With these conditions, lifting the boundary layer parcel having the lowest value of \( \theta_{ep} \) (that parcel just below the capping inversion, in this instance) yields a CAPE value of \( \sim3500 \) J kg\(^{-1}\) (70% of the MLCAPE). This is approximately the lowest amount of CAPE that a boundary layer parcel being ingested into the updraft would have available to it and is, thus, quite large. Therefore, for this sounding, if RFD parcels reaching the ground generally originated from the boundary layer, such air parcels would have significant potential buoyancy despite the magnitude of \( \Delta \theta_{ep} \) in this case.

Because small (more positive) \( \Delta \theta_{ep} \) values are consistent with relatively constant CAPE values for air parcels resident with the boundary layer, another possible measure of tornado potential is the minimum CAPE that can be achieved for any air parcel originating from the boundary layer. Thus, the CAPE for the lowest-\( \theta_{ep} \) parcel from the boundary layer was calculated for the database of soundings presented in section 3 and a box-and-whiskers diagram was developed to compare categories (Fig. 12). It is clear that significant tornado events had greater CAPE values, although significant overlap is present across categories, and the difference in means is only significant at the 85% confidence interval (\( \alpha = 0.15 \)). Interestingly, the lowest boundary layer \( \theta_{ep} \) CAPE from the 0000 9 May 2003 KOUN sounding is well over the mean CAPE value calculated in Fig. 12 (1598 J kg\(^{-1}\)).

Two significant tornado cases from the sounding sample used in section 3 also had boundary layer \( \Delta \theta_{ep} \) values approaching \( \sim10 \) K, and thus are similar to the 0000 UTC 9 May 2003 KOUN sounding in that they are outliers. One of these tornadic cases was associated with a supercell, as was the 9 May 2003 event. Both events had MLCAPE values in excess of 2000 J kg\(^{-1}\) as well as

### Table 2. A summary of selected boundary layer parameters from eight observed soundings taken in the inflow region of storms that produced violent (F4 and F5) tornadoes. The soundings were collected within 115 km and 75 min of the violent tornadoes.

<table>
<thead>
<tr>
<th>Sounding date and location</th>
<th>Max tornado intensity</th>
<th>Surface ( \theta_{ep} ) (K)</th>
<th>Mean boundary layer temperature lapse rate (°C km(^{-1}))</th>
<th>Mean boundary layer relative humidity (%)</th>
<th>Boundary layer ( \Delta \theta_{ep} ) (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0000 UTC 9 Jun 1966, Topeka, KS</td>
<td>F5</td>
<td>345.3</td>
<td>4.1</td>
<td>91</td>
<td>7.6</td>
</tr>
<tr>
<td>0000 UTC 1 Jun 1985, Pittsburgh, PA</td>
<td>F5</td>
<td>344.6</td>
<td>7.7</td>
<td>82</td>
<td>1.1</td>
</tr>
<tr>
<td>0000 UTC 14 May 1995, Lincoln, IL</td>
<td>F4</td>
<td>344.8</td>
<td>8.5</td>
<td>96</td>
<td>–1.8</td>
</tr>
<tr>
<td>VORTEX1 8 Jun 1995, Allison, TX</td>
<td>F4</td>
<td>364.1</td>
<td>6.9</td>
<td>84</td>
<td>3.2</td>
</tr>
<tr>
<td>0000 UTC 15 Apr 1996, Little Rock, AR</td>
<td>F4</td>
<td>332.4</td>
<td>8.1</td>
<td>92</td>
<td>3.7</td>
</tr>
<tr>
<td>0000 UTC 4 May 1999, Norman, OK</td>
<td>F5</td>
<td>343.6</td>
<td>6.5</td>
<td>86</td>
<td>–0.3</td>
</tr>
<tr>
<td>0000 UTC 9 May 2003, Norman, OK</td>
<td>F4</td>
<td>364.2</td>
<td>8.5</td>
<td>81</td>
<td>–5.1</td>
</tr>
<tr>
<td>0000 UTC 6 Feb, 2008 Little Rock, AR</td>
<td>F4</td>
<td>337.3</td>
<td>7.7</td>
<td>85</td>
<td>–0.7</td>
</tr>
<tr>
<td>Mean</td>
<td>—</td>
<td>347.0</td>
<td>7.3</td>
<td>87</td>
<td>1.0</td>
</tr>
<tr>
<td>Median</td>
<td>—</td>
<td>344.7</td>
<td>7.7</td>
<td>86</td>
<td>0.4</td>
</tr>
</tbody>
</table>
0–1-km bulk shear values of 15 m s$^{-1}$ and 0–1-km SRH of 250 m$^2$ s$^{-2}$ or more. Thus, it appears that even in the presence of strongly negative $\Delta \theta_{ep}$, some potential for significant tornadoes still exists. Given the cases discussed here, this may be especially true in events characterized by both extreme instability and very large values of low-level shear or SRH. The CAPE values computed using the lowest-$\theta_{ep}$ parcels in the boundary layers were $\sim$1280 and 1750 J kg$^{-1}$ for these two cases, respectively, which are approximately 62% and 79% of the MLCAPE in these two soundings. This is above the mean CAPE value for significant tornadoes in Fig. 12, although the values are not as large as with the 0000 9 May 2003 KOUN case. Nonetheless, lowest boundary layer $\theta_{ep}$ CAPE may be worthy of some consideration, especially for large CAPE and shear environments that are also characterized by large (more negative) values of boundary layer $\Delta \theta_{ep}$.

To provide additional examples of boundary layer $\Delta \theta_{ep}$, two cases not used in the development of this study’s statistical results are briefly analyzed in the following subsections. In both cases, the 0–1-km bulk shear and SRH were favorable for significant tornadoes based on T03 and results obtained herein. Except for the NSE of the first storm discussed in case two, MLLCL heights were also well within the ranges favorable for significant tornadoes.

1) NORTH-CENTRAL OKLAHOMA—10 MAY 2010

A significant tornado with a pathlength of 66 km and maximum damage rating of EF3 impacted the area from near Wakita, Oklahoma, to South Haven, Kansas, between 2038 and 2134 UTC on 10 May 2010. An observed sounding was taken from Lamont, Oklahoma (KLMN), at 2100 UTC 10 May 2010 (Fig. 13). KLMN is approximately 37 km southeast of the EF3 damage that occurred near Wakita. The KLMN sounding had $\sim$1300 J kg$^{-1}$ of MLCAPE and $\sim$40 m s$^{-1}$ of 0–6-km bulk shear. The 0–1-km bulk shear was 20 m s$^{-1}$, and the 0–1-km SRH, based on the observed storm motion, was $\sim$625 m$^2$ s$^{-2}$. The MLLCL height was 775 m AGL. In addition, the boundary layer $\Delta \theta_{ep}$ value was $\sim$1.2 K, which is well within the favored range of boundary layer $\Delta \theta_{ep}$ values for the significant tornadoes in both Figs. 6 and 9.

2) WESTERN NEBRASKA—24 MAY 2010

A supercell occurred within a “break” in a line of MC storms 35 km west-northwest of North Platte, Nebraska (KLBF), around 2300 UTC 25 May 2011. No tornadoes were reported, despite both operator- and algorithm-defined tornadic vortex signatures (TVSs). The observed 0000 UTC KLBF sounding shown in Fig. 14 contained 2600 J kg$^{-1}$ of MLCAPE and $\sim$27 m s$^{-1}$ of 0–6-km bulk shear. The 0–1-km bulk shear was 18 m s$^{-1}$, and
observed storm motion yielded 0–1-km SRH of \( \sim 300 \text{ m}^2 \text{ s}^{-2} \). The MLLCL height was \( \sim 1200 \text{ m AGL} \), which is marginal for significant tornadoes based on T03 and the present study. However, boundary layer \( \Delta \theta_{ep} \) was more clearly outside of the ranges favorable for significant tornadoes. If the minor cap at 801 hPa is considered to be the top of the boundary layer since it fits the definition of a cap used to develop the statistical results in this study, then \( \Delta \theta_{ep} \) is \(-4.5 \text{ K}\). Using the more distinct cap at 730 hPa as the top of the boundary layer yields an even larger \( \Delta \theta_{ep} \) value of \(-16.9 \text{ K}\).

Several nontornadic DSs with strong low-level rotation and TVSs also occurred in close temporal and spatial proximity to an observed 1800 UTC 24 May 2010 KLBF sounding (not shown). The 0–1-km bulk shear was \( \sim 17 \text{ m} \text{s}^{-1} \) and 0–1-km SRH, based on observed storm motion, was \( \sim 150 \text{ m}^2 \text{ s}^{-2} \). The MLLCL height was also only 660 m AGL, but boundary layer \( \Delta \theta_{ep} \) was \(-7.5 \text{ K}\), which is well outside of the favorable range for significant tornadoes.

### b. Relation to LCL and LFC height

Correlation coefficients were calculated using the dataset presented in section 3 to determine if there is some relationship between boundary layer \( \Delta \theta_{ep} \) and MLLCL or MLLFC height in order to determine if the indices are providing the same information. The resultant coefficient matrix in Table 3 suggests that there is a correlation at the 95% confidence interval \((\alpha = 0.05)\) only for the categories of events that include significant tornadoes. The strongest correlation \((r = 0.77)\) was between boundary layer \( \Delta \theta_{ep} \) and MLLCL height for the supercell-only significant tornado cases. These two parameters could be correlated because they provide an indication of the same information, or they could be correlated, but not in a cause-and-effect type of manner. Though it has not been tested here, operational experience suggests that the latter may be more likely as the parameters can easily vary independently of each other. For instance, it is not difficult to envision a low-level sounding profile that has a favorably low MLLCL height, but a significantly negative (and thus unfavorable) boundary layer \( \Delta \theta_{ep} \) value resulting from a sharp decrease in moisture within the boundary layer, but primarily above the 100-hPa mixed layer. Alternatively, either low or high LCL heights could exist with a boundary layer \( \Delta \theta_{ep} \) value of 0 K in an environment containing a dry-adiabatic lapse rate and a constant water vapor mixing ratio with height. Simply moving the mixing ratio value to the left or right on the resulting sounding would produce drastically different LCL heights even with the same boundary layer \( \Delta \theta_{ep} \) of 0 K. A common example of this scenario is the appearance of a deeply mixed boundary layer possessing an “inverted v” structure and, thus, minimal boundary layer \( \Delta \theta_{ep} \), but a high LCL height. Such an environment often portends high-based convection with a reduced tornado threat.

Despite the fact that boundary layer \( \Delta \theta_{ep} \) and LCL height may not be related to one another in a direct manner, the fact that they show a tendency to be correlated for significant tornadoes (especially those associated with supercells) indicates that they are at least in some way providing the same information. For nontornadic events, boundary layer \( \Delta \theta_{ep} \) and LCL height seem to be relatively independent of each other, which is where both can provide significant value (as opposed to one or the other).

### c. Application to operational forecasting

It appears that boundary layer \( \Delta \theta_{ep} \) provides enough discrimination between nontornadic and significantly tornadic NSEs to provide useful information to operational forecasters. To further quantify its usefulness, both the dataset in section 3 and the one from T07 were used to calculate the probability of detection (POD), false alarm rate (FAR), frequency of correct null forecasts (FOCN), critical success index (CSI), and true skill score (TSS) for the significant tornado events versus the nontornadic events (both with and without surface-based stable layers) using \( 2 \times 2 \) contingency tables as in

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**Figure 12.** As in Fig. 5, but for CAPE (J kg\(^{-1}\)) calculated by lifting the parcel in the boundary layer having the lowest \( \theta_{ep} \) value.
Doswell et al. (1990). A summary of these statistical measures for the range of boundary layer $\Delta \theta_{ep}$ values is presented in Fig. 15. The most prominent difference between the dataset presented in section 3 and T07’s dataset is that the FAR is consistently higher in the T07 results shown in Fig. 15b. Even so, the FOCN is also consistently high in both sets of results.

Based on the results in Fig. 15a, forecast skill is maximized for a boundary layer $\Delta \theta_{ep}$ value of $-3$ K, where the POD is 0.86, the FAR is 0.39, the FOCN is 0.88, the CSI is 0.55, and the TSS is 0.53. In the T07 sounding sample illustrated in Fig. 15b, the POD for a boundary layer $\Delta \theta_{ep}$ value of $-3$ K is 0.70, and the FAR is much larger at 0.74. Nonetheless, a boundary layer $\Delta \theta_{ep}$ value of $-3$ K appears to be a reasonable guideline for discriminating between significantly tornadic and nontornadic NSEs.

POD, FAR, and TSS were also calculated for the MLLCL heights of the data presented in section 3, and from that it was determined that the best discriminating MLLCL value (at least using cases from the data in section 3) is approximately 1000 m AGL. Contingency tables for the significant tornado events versus the nontornadic events using a boundary layer $\Delta \theta_{ep}$ value of $-3$ K and a MLLCL height of 1000 m AGL as discriminators were also developed using the data from section 3 to compare the relative performance between the two (Table 4). The results in Table 4 suggest that the skill associated with the two parameters is nearly equal. An MLLCL height of 1000 m AGL did have a slightly higher POD (0.90) than a boundary layer $\Delta \theta_{ep}$ value of $-3$ K (0.86), but that came at the expense of a slightly higher FAR. However, the purpose of this study is not to suggest the sole use of boundary layer $\Delta \theta_{ep}$ in characterizing how favorable a thermodynamic NSE may be for tornadogenesis, but rather to offer another tool for forecasters to use. In addition, the intent is not to suggest reliance on specific numbers for determining tornado threat, but rather to provide application of modeling results in a manner that is easy to visualize and compare to other well-known parameters. It is hoped that the apparent utility of boundary layer $\Delta \theta_{ep}$ encourages forecasters to more closely inspect the low-level $\theta_{ep}$ profile of soundings in tornadic and nontornadic NSEs, while remaining mindful of physical processes that are relevant to tornadogenesis. It is noted that while the results herein are supportive of the hypothesis posed by A04 that downdrafts are significantly inhibited by caps and, thus, that boundary layer $\theta_{ep}$ may be important to tornadogenesis, these results do not constitute validation of that hypothesis. In fact, it is possible that other physical mechanisms, including ones involving the updraft,
could be responsible for the $\Delta \theta_{cp}$ differences observed herein. Verifying the physical mechanisms responsible for these observed $\Delta \theta_{cp}$ differences will require further testing.

d. Potential failure modes

The minimal overlap shown between significant tornado cases and nontornadic events for soundings without surface-based stable layers using the data presented in section 3 suggests that boundary layer $\Delta \theta_{cp}$ may be useful in characterizing potential RFD buoyancy and tornado potential, given the apparent relation between below-cap $\theta_{cp}$ profiles to low-level RFD $\theta_{cp}$ deficits (A04). However, the processes leading to the formation and intensification of the RFD are no doubt complex and, thus, the potential buoyancy of the RFD is not always easy to anticipate. For this reason, there are situations where boundary layer $\Delta \theta_{cp}$ may not accurately reflect the true tornado potential. For instance, under the right conditions (e.g., very heavy precipitation), relatively high-altitude air having low $\theta_{cp}$ can descend to the surface, even in the presence of a capping inversion and high boundary layer $\theta_{cp}$ values (A04). This may be especially common in MC cases, which may be why boundary layer $\Delta \theta_{cp}$ was a poor predictor for the MC events in this study. In addition, the nontornadic supercells in the section 3 data that were subjectively classified as being high precipitation based on radar reflectivity characteristics contained higher (more positive) mean boundary layer $\Delta \theta_{cp}$ values than classic supercells (means of $-3.3$ and $-8.3$ K, respectively). This may imply that high-precipitation supercells do indeed drive lower $\theta_{cp}$ to the ground and, thus, may require particularly favorable boundary layer $\theta_{cp}$ profiles for tornadogenesis (the mean $\Delta \theta_{cp}$ value for the high-precipitation supercells that did produce significant tornadoes was $1.4$ K, compared to $-1.6$ K for the classic supercells).

![FIG. 14. As in Fig. 2, but for the 0000 UTC 25 May 2010 KLBF observed sounding.](image)

<table>
<thead>
<tr>
<th>Correlation with boundary layer $\Delta \theta_{cp}$</th>
<th>Coefficient for MLLCL height</th>
<th>Coefficient for MLLFC height</th>
</tr>
</thead>
<tbody>
<tr>
<td>All cases</td>
<td>0.49</td>
<td>0.18</td>
</tr>
<tr>
<td>Nontornadic cases</td>
<td>0.29</td>
<td>0.05</td>
</tr>
<tr>
<td>Significant tornado cases</td>
<td>0.65</td>
<td>0.32</td>
</tr>
<tr>
<td>All supercell cases</td>
<td>0.56</td>
<td>0.21</td>
</tr>
<tr>
<td>Supercell-only nontornadic cases</td>
<td>0.30</td>
<td>0.06</td>
</tr>
<tr>
<td>Supercell-only significant tornado cases</td>
<td>0.77</td>
<td>0.20</td>
</tr>
</tbody>
</table>
The distribution of $\theta_{ep}$ within the boundary layer may also be important. For instance, a sounding could contain a thin dry layer just below the base of the cap and at the top of the boundary layer that could result in a very low boundary layer $\Delta \theta_{ep}$ value. However, owing to entrainment as the RFD descends, the actual $\theta_{ep}$ of the RFD when it reaches the surface may be much higher than that suggested by boundary layer $\Delta \theta_{ep}$. Thus, while boundary layer $\Delta \theta_{ep}$ provides a simple framework from which to help determine RFD characteristics, it cannot provide all of the information associated with physical processes affecting the RFD that might be relevant to tornado development. This reiterates the need to thoroughly inspect individual soundings such that the use of a measure like $\Delta \theta_{ep}$ does not obfuscate the physical processes that lead to tornadogenesis failure or success.

### e. Limitations and future work

The RUC-based analyses used in this study may not adequately resolve mesoscale and smaller-scale thermodynamic structures, which is a potential limitation of this study. Rapid and significant changes in low-level sounding profiles routinely occur. While every attempt was made to use the analysis that appeared to most adequately represent each NSE, 1-h snapshots of conditions almost certainly provide imperfect representations of thermodynamic structures.

Future research should investigate the spatial and temporal tendencies of boundary layer $\Delta \theta_{ep}$, as well as the distribution of $\theta_{ep}$ within the boundary layer in particular cases. Further analysis of high-CAPE or high-shear cases in which tornadogenesis occurred despite low values of boundary layer $\Delta \theta_{ep}$ would also be useful. Given the results in Fig. 7, it may also be worthwhile to explore the operational usefulness of a parameter combining boundary layer $\Delta \theta_{ep}$ and low-level shear or SRH. Finally, given the importance hydrometeor forcing likely plays in determining RFD character (Kumjian 2011), future research should investigate the impacts variations in hydrometeor properties have on the utility
of boundary layer $\Delta \theta_{ep}$, since such variations (e.g., stronger downdraft forcing) could theoretically contribute to significant descent of relatively low $\theta_{ep}$ air in the RFD to the ground, even if the presence of a cap suggests that this would not occur.

5. Summary and conclusions

It is relatively common for strongly rotating supercells to fail to produce significant tornadoes, even when most NSE parameters indicate that they should occur. M02 concluded that relatively warm, moist, and potentially buoyant RFDs are supportive of tornadogenesis. “Cold” RFDs, on the other hand, reduce the potential for tornadogenesis (M02). Based upon the findings of Askelson (2002) and A04, the hypothesis that $\theta_{ep}$ profiles within capped boundary layers are important to tornadogenesis has been studied.

The following points summarize the key findings of this study (for supercell tornadoes):

- Boundary layer $\Delta \theta_{ep}$ appears to have appreciable skill in discriminating between nontornadic and significantly tornadic NSEs. Values of boundary layer $\Delta \theta_{ep}$ greater than $-3$ K are especially favorable for the development of significant tornadoes. Several NSEs that were examined in this study and that were associated with violent tornadoes had boundary layer $\Delta \theta_{ep}$ values of 0 K or greater.
- The skill of boundary layer $\Delta \theta_{ep}$ appears greatest with DSs. It may be a less reliable predictor with MC storms, and in cases with surface-based stable layers.
- The performance of boundary layer $\Delta \theta_{ep}$ in discriminating nontornadic and significantly tornadic NSEs, and its development based upon previous modeling results, suggests that an assessment of the vertical $\theta_{ep}$ profiles may indeed provide a means by which forecasters can attempt to estimate low-level RFD characteristics using a physically based approach.

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REFERENCES


