

**Observations of a squall line and its near environment using high-frequency  
rawinsonde launches during VORTEX2**

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## Abstract

Rawinsonde data were collected before and during passage of a squall line in Oklahoma on 15 May 2009 during VORTEX2. Nine soundings were released within 3 hours, allowing for unprecedented analysis of the squall line’s internal structure and nearby environment. Four soundings were released in the pre-storm environment and they document the following features: low-level cooling associated with the reduction of solar isolation by a cirrus anvil; abrupt warming (1.5 K in 30 min) above the boundary layer, which is probably attributable to a gravity wave; increases in both low-level and deep-layer vertical wind shear within 100 km of the squall line; and evidence of ascent extending at least 75 km ahead of the squall line. The next sounding was released  $\sim 5$  km ahead of the squall line’s gust front; it documented a moist absolutely unstable layer within a 2-km deep layer of ascent, with vertical air velocity of approximately  $6 \text{ m s}^{-1}$ . Another sounding was released after the gust front passed but before precipitation began; this sounding showed the cold pool to be  $\sim 4$  km deep, with a cold pool intensity  $C \approx 35 \text{ m s}^{-1}$ , even though this sounding was located only 8 km behind the surface gust front. The final three soundings were released in the trailing stratiform region of the squall line, and they showed typical features such as: “onion” shaped soundings; nearly uniform equivalent potential temperature over a deep layer; and an elevated rear inflow jet. The cold pool was 4.7 km deep in the trailing stratiform region, and extended  $\sim 1$  km above the melting level, suggesting that sublimation was a contributor to cold pool development. A mesoscale analysis of the sounding data shows an upshear tilt to the squall line, which is consistent with the cold pool intensity ( $C$ ) being much larger than a measure of environmental vertical wind shear ( $\Delta U$ ). This dataset should be useful for evaluating cloud-scale numerical model simulations and analytic theory, but the authors argue that additional observations of this type should be collected in future field projects.

# 1 Introduction

Squall lines are linearly organized thunderstorm complexes that often produce severe wind gusts, heavy rain, and lightning. In recent decades, much has been learned about squall line dynamics from Doppler radar data and through numerical simulation (see, e.g., Johnson and Mapes 2001; Fritsch and Forbes 2001). However, in-situ observations of temperature and moisture have been relatively rare, particularly in mid-latitude squall lines because operational rawinsonde networks are too coarse to resolve mesoscale structure, and because airplane flights in the convective region are too hazardous. This article presents unique analyses of a squall line and its nearby environment using high-frequency rawinsonde launches from ground-based mobile platforms.

Before proceeding, we review briefly the important thermodynamic features of squall lines, and techniques that have been used to observe them. It has long been recognized that relatively cold air arrives at the surface during squall line passage [see, e.g., the review by Newton (1950)]. Meteorologists usually refer to the near-surface cold air in squall lines simply as “cold pools,” although the terms “density currents” and “gravity currents” are also used frequently. Cold pools are easily observed from surface data. A thorough study was published recently by Engerer et al. (2008) using high-temporal-resolution observations from the Oklahoma Mesonet during the warm season. Engerer et al. (2008) found that mature mesoscale convective systems were accompanied by average temperature drops of  $\sim 7^\circ\text{C}$  and average pressure increases of  $\sim 4.5$  mb. They noted (p. 4846) that the vertical extent of cold pools remains largely unknown.

Rawinsonde data are usually used to deduce cold pool properties above the surface. However, high temporal resolution (of order 1 hour) or spatial resolution (of order 10 km) is necessary to document conditions before and after passage of the surface gust front. A summary of observed cold pool properties from rawinsonde observations was provided by Weisman and Rotunno (2005) (their Table 1). They found cold pool depths up to 2.5 km and cold pool intensity (as measured by the parameter  $C$ , explained later) up to  $30\text{ m s}^{-1}$ .

However, some significantly deeper cold pools have been documented. For example, Roux (1988) documented a cold pool depth of  $\sim 4$  km in a squall line over west Africa. In this case, the cold pool sounding was taken “in the rear part of the trailing stratiform region” (p. 407), leading one to question whether the squall-line cold pool was perhaps shallower at the leading edge of the system. In the central United States, Wakimoto (1982) used rawinsonde data from project NIMROD and found one case with a cold pool depth of 3.95 km. He noted that this depth is “a much larger value than is normally associated with the gust front.”

All of the analyses cited above used two soundings to document the cold pool: one in the cold pool itself (hereafter referred to as a “cold pool sounding”), and one ahead of the system (hereafter referred to as an “environmental sounding”). More comprehensive analyses utilizing soundings at multiple locations are quite rare. To our knowledge, the highest-resolution direct thermodynamic measurements of a squall line cold pool using rawinsonde data (prior to this article) were obtained more than 60 years ago during The Thunderstorm Project in southwest Ohio (Byers and Braham 1949). On 29 May 1947, a squall line passed the project’s array of six closely-spaced rawinsonde sites near Wilmington, Ohio. Some sites released more than one rawinsonde, and so a total of 10 soundings were obtained: 2 just before passage, and 8 within 3.4 hours after squall line passage. The data were originally analyzed and reported by Newton (1950). Further analysis of this dataset was presented in Newton and Newton (1959). The analysis remains a classic in squall-line observations, and is often reproduced in review articles on squall lines (e.g., Newton 1963; Hane 1986). Vertical cross sections of temperature and potential temperature [see Figs. 10–11 in Newton (1950)] show that the cold pool was  $\sim 3$  km deep. The analysis of cross-line system-relative wind [see Fig. 14 in Newton (1963)] shows an elevated rear-inflow jet near the top of the cold pool.

Newton (1950) was perhaps the first to determine that air in squall line cold pools originates in mid levels ( $\sim 5$  km AGL) and has very low wet-bulb potential temperature ( $\theta_w$ ). A mesoscale, sub-saturated downdraft toward the back of the squall line (at least 100 km

behind the surface gust front) was responsible for the transport of this low- $\theta_w$  air from midlevels to low-levels in this case. Evaporation of rain was inferred to be the mechanism that cooled the air. These processes have since been supported by numerous observational studies and numerical simulations.

Several times during the 1960s and 1970s, the National Severe Storms Laboratory conducted special field experiments in which rawinsondes were released from multiple sites in Oklahoma with relatively high frequency ( $\sim 3$  h). Using these data, mesoscale analyses of squall line events were published by Sanders and Paine (1975), Sanders and Emanuel (1977), Ogura and Chen (1977), Ogura and Liou (1980), and Park and Sikdar (1982). Broadly speaking, these studies found similar features to Newton (1950). In the analysis of Ogura and Liou (1980), the cold pool was  $\sim 4$  km deep, according to their Fig. 21. Based on the potential temperature analysis of Sanders and Paine (1975) (their Fig. 10), their case had a cold pool  $\sim 3$  km deep.

More recently, rawinsonde data were collected using mobile platforms during the Bow Echo and MCV Experiment (BAMEX; Davis et al. 2004). The BAMEX dataset included mobile surface-based radiosondes as well as dropsondes from a jet flying at  $\sim 12$  km. Several mesoscale convective systems were sampled. Analysis of cold pool properties by Bryan et al. (2004, 2005) and Ahijevych et al. (2006) showed that cold pools were occasionally greater than 4 km deep, even at the leading edge of the system (Bryan et al. 2005).

Despite several analyses that show cold pool depths up to 4 km, cold pool depths are generally quoted in the literature to be about 1-2 km deep (a factor of two smaller). These smaller depths seem to be true for most unorganized convection (e.g., Mahoney 1988), supercellular thunderstorms, and tropical oceanic MCSs [see, e.g., review in Chapters 8 and 9 of Houze (1993)]. Many idealized simulations of midlatitude-type convective systems also produce the 1–2 km depth (e.g., Rotunno et al. 1988; Weisman 1992; Weisman and Rotunno 2004; Parker and Johnson 2004), although many early simulations neglected ice microphysics and used highly idealized initial conditions. Regardless of the origin for the

1–2 km depth, it seems that relatively few analyses of mature mid-latitude MCSs have been conducted/published using direct thermodynamic observations.

In addition to uncertainties about the properties of squall line cold pools, some recent studies have shown that squall lines can modify their environment in significant ways. Most of these studies have utilized numerical simulation or analytic theory. For example, Nicholls et al. (1991), Mapes (1993), Fovell (2002), and Fovell et al. (2006) have shown how gravity waves triggered by convective heating can modify stability, humidity, and vertical wind shear in the nearby environment (i.e., within about 100 km). As far as we can tell, there are few observational studies that have evaluated these modeling/theoretical studies.

In this article, we analyze a unique dataset collected during the Second Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX2). The primary focus of this field project was to study supercellular thunderstorms and tornadoes. However, on 15 May 2009 a squall line passed over the VORTEX2 armada of mobile observing facilities. Herein, we present an analysis of the mobile rawinsonde data, primarily using 9 rawinsonde launches over a 3-hour time period. The following analyses support the earlier observational studies (discussed above) showing that MCS cold pools can be 4 km deep in mature mid-latitude MCSs. The soundings launched before passage of the squall line also show the evolution of the near-squall-line environment with unprecedented detail.

## **2 Overview of the 15 May 2009 squall line**

The 15 May 2009 squall line formed at approximately 2030 UTC along a cold front that extended from the northern Texas panhandle to southern Minnesota. Behind the front, temperature and moisture steadily decreased to the north and west (Fig. 1). Ahead of the front in Oklahoma, temperature and moisture were mostly uniform with temperatures around 30 °C and dewpoint temperatures around 20 °C. The front was also marked by strong surface convergence, with 15–20 kt southerly flow to its south and 15–20 kt northerly flow

to its north.

On 15–16 May, field observations for VORTEX2 were collected primarily near Cherokee, Oklahoma, within Alfalfa County (shown in inset of Fig. 1) from approximately 2130 UTC 15 May to 0100 UTC 16 May 2009. Rawinsonde launches during this time period are summarized in Table 1. At the beginning of this time period, the squall line was still becoming organized but was characterized by almost continuous reflectivity  $> 40$  dBZ from the northern Texas panhandle to northeastern Kansas (Fig. 2a). Severe hail was reported in several locations of northwest Oklahoma and south-central Kansas in the early stages of squall line development. Soundings S1–S3 were released well ahead of the squall line ( $> 40$  km).

By 2305 UTC, the squall line gust front was nearing VORTEX2 facilities. WSR-88D data from KVNK at this time (Fig. 3a) show a continuous line of strong reflectivity ( $> 45$  dBZ) from the southwest to northeast near the Kansas/Oklahoma border. A persistent minimum in reflectivity to the west-southwest of KVNK is attributable to partial beam blockage, and is denoted in Fig. 3 by dashed lines. Two soundings (S4 and S5) were released at this time, at locations shown on Fig. 3a. A vertical cross section of reflectivity is shown in Fig. 4a; the location of this cross section is shown by a dashed gray line in Fig. 3a. For vertical cross sections, data from KVNK were interpolated onto a Cartesian grid with 1 km grid spacing using a Barnes (1973) analysis (radius of influence = 2 km; non-dimensional smoothing parameter = 0.5). To account for system movement during the volume scan, data are advected to a common time assuming a storm motion of  $16 \text{ m s}^{-1}$  to the southeast. At this time, the high-reflectivity core was roughly upright, and a trailing stratiform region was apparent above the melting layer (Fig. 4a).

The next sounding was released 32 minutes later (S6 in Fig. 3b). This sounding was launched after the gust front passed but before rain began. A fine line, denoting the gust front, can be seen in Fig. 3b. The squall line’s trailing stratiform region was becoming notable in the lowest elevation scan of KVNK (Fig. 3b) at this time. A vertical cross section

(Fig. 4b) shows that the trailing stratiform precipitation was beginning to reach the ground over an area roughly 40-km behind the convective region.

The next sounding was released 37 minutes later in the trailing stratiform region (S7 in Fig. 3c). Because of the beam blockage, the trailing stratiform region is not obvious in Fig. 3c, although light rain was falling at the surface at this time. The structure of the trailing stratiform region is more obvious in the vertical cross section (Fig. 4c), which shows  $>25$  dBZ below the melting level extending more than 50 km behind the convective region. Two more soundings were released in the trailing stratiform region within the next 30 minutes (Table 1).

By the end of the VORTEX2 observing period, the convective region of the squall line had moved into north-central Oklahoma and developed a bow shape to the east-southeast of the primary sounding site (Fig. 2b). Widespread severe winds were reported along the Kansas/Oklahoma border, near the northern end of the bow. No severe weather was reported in Alfalfa County during this event.

Surface observations from the Oklahoma Mesonet station in Cherokee, Oklahoma, are shown in Fig. 5. Three distinct time periods are apparent in these observations, as delineated by the vertical gray lines in Fig. 5. Before 2150 UTC 15 May, the temperature, water vapor mixing ratio, and wind speed were all roughly constant. The pressure dropped  $\sim 2$  mb in 45 minutes starting at  $\sim 2100$  UTC, perhaps because of the approaching cold front. The first VORTEX2 sounding was released at 2138 UTC, near the end of this pressure drop. (Note that sounding launches are denoted by large dots in panels a–c of Fig. 5.)

At 2150 UTC, an optically thick cirrus anvil cloud began blocking the sun at Cherokee. This fact is apparent in the sudden drop in incoming solar radiation at the Oklahoma Mesonet site (Fig. 5e). A visible satellite image from 2139 UTC (Fig. 6) shows the shadow just to the west of Cherokee. The air temperature at Cherokee then decreased steadily, and  $4^\circ\text{C}$  of cooling occurred over 90 minutes (Fig. 5a). This amount of cooling is consistent with observations and modeling of anvil shadows from deep cumulonimbus clouds in Oklahoma at

this time of year (Markowski et al. 1998; Markowski and Harrington 2005). The water vapor mixing ratio and wind direction were approximately constant in this time period (Fig. 5c–d) which suggests that Cherokee was experiencing the same airmass (i.e., pre-frontal).

The pressure time-series was complex between 2150 UTC and 2300 UTC, with a 1 mb pressure rise immediately after anvil shading began, followed by a 1 mb pressure drop, and then several mb of pressure rise, all during a 70 minute time period. The sharp pressure rise that started at 2150 UTC might be related to the anvil shading; that is, surface pressure should have increased in hydrostatic response to the near-surface cooling. The sharp pressure drop that started at 2210 UTC might be associated with a propagating gravity wave, perhaps triggered by the formation of the squall line in Texas or Kansas. Using high-temporal-resolution surface observations, Adams-Selin and Johnson (2009) documented a pressure drop of similar amplitude during formation of a bow echo in Oklahoma. These explanations for the surface pressure changes during this time period are evaluated further using VORTEX2 data later in this article.

The surface gust front of the squall line passed Cherokee at  $\sim 2320$  UTC. Temperature and water vapor mixing ratio both decreased rapidly and surface pressure increased sharply. The maximum wind gust ( $16.3 \text{ m s}^{-1}$ ) occurred shortly after gust front passage but before precipitation started. At the primary sounding site ( $\sim 10$  km to the south of Cherokee), heavy precipitation began shortly after 2340 UTC, or roughly 15 minutes after passage of the gust front. The wind speed at Cherokee dropped significantly as precipitation began.

After 0000 UTC 16 May, conditions at Cherokee were more quiescent. An exception was a sharp pressure drop of  $\sim 2$  mb starting at 0100 UTC, followed by a sharp pressure rise of equal amplitude. This feature might have been associated with a wake low (e.g., Johnson and Hamilton 1988) at the trailing edge of the stratiform precipitation region.

Based on Oklahoma Mesonet observations and wind profiler data (not shown), the cold front likely passed Cherokee at about 0300 UTC 16 May. The final VORTEX2 sounding was launched two hours earlier, and so the cold front was not sampled by rawinsonde data

presented herein.

### 3 Rawinsonde data

Nine soundings were released in Alfalfa County as part of special data collection during VORTEX2. All but one of the soundings were released from a location 10 km south of Cherokee, Oklahoma (see inset of Fig. 1). The times of these launches, and conditions at launch time, are listed in Table 1. One sounding (S5) was released from a location 6 km northwest of Cherokee at the same time as S4 (see Table 1). All soundings used Vaisala RS92 radiosondes using the Mobile GPS Advanced Upper-Air Sounding System (MGAUS) developed at the National Center for Atmospheric Research (NCAR). Quality control of the rawinsonde data, including bias correction, was performed by staff at NCAR’s Earth Observing Laboratory (EOL).<sup>2</sup> Erroneous measurements caused by wetting and/or icing of the sensors are difficult to correct and are not accounted for herein, although later we point out some instances where such errors may have occurred.

In Table 1, the time of sounding launch relative to gust-front passage (third column) was calculated assuming the gust front passed the primary sounding site at 2328 UTC and assuming it passed the location of S5 at 2311 UTC. During the time period of rawinsonde launches, the squall line’s convective region moved to the southeast at an average speed of  $16 \text{ m s}^{-1}$ . The approximate distance of the soundings from the squall line’s gust front at launch (fourth column in Table 1) is calculated using time-to-space conversion with this average motion. The locations of *all* sounding data are shown in Fig. 7, where  $x_c$  is the estimated across-line distance (based on constant system propagation of  $16 \text{ m s}^{-1}$ ). S6 had missing thermodynamic information above 350 mb (note gray dots in Fig. 7), although winds were obtained from GPS tracking up to 150 mb.

Relative to the squall line, the soundings appear to tilt toward the rear of the squall line

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<sup>2</sup>Details are available in a 9-page document provided with this dataset at <http://data.eol.ucar.edu>; the document is also available from NCAR/EOL upon request.

(Fig. 7); the lone exception is S6 at low levels. These system-relative sonde trajectories show the combined effects of the moving squall line and advection of the sondes by the winds, the net effect being that the sonde motion is towards the rear of the squall line at nearly all locations. This issue is raised here so that the following skew-T plots can be interpreted properly. Specifically, readers should be aware that the measurements are not in a upright column. For pre-squall-line (environmental) soundings, the low-level data are farther away from the squall line than upper-level data.

For reference, we plot surface data from the rawinsondes as dots on Figs. 5a–c. For reasons that are unclear, the temperature, pressure, and moisture data were notably different from values at the Cherokee mesonet site. Specifically, in the rawinsonde’s surface data, temperature was  $\sim 2$  K higher, pressure was  $\sim 1.8$  mb lower, and water vapor mixing ratio was  $\sim 0.5$  g kg $^{-1}$  higher than data from the Cherokee mesonet site. The differences may be attributable to one or several of the following: different land-surface conditions at the two sites (which are separated by  $\sim 10$  km); different elevations of the two sites (Cherokee is  $\sim 15$  m lower than the primary sounding site); and/or different observation heights (the Oklahoma mesonet observations were at 1.5 m AGL, whereas the radiosondes were held at  $\sim 1$  m AGL just before launch). Nevertheless, the trends in time are consistent between the two datasets, and so the processes inferred from the Cherokee mesonet data (in the previous section) were clearly captured by the near-surface rawinsonde data.

### **a. Pre-squall-line environment**

The first sounding (S1, Fig. 8a) was released in clear-sky conditions. It showed a 1.2-km deep well-mixed boundary layer, and a shallow super-adiabatic layer near the surface. There was a weak capping inversion above the boundary layer. Farther aloft there were multiple elevated mixed layers. Using a 500-m mixed-layer source parcel, convective available potential energy (CAPE) was 4260 J kg $^{-1}$  and convective inhibition (CIN) was 8 J kg $^{-1}$  (Fig. 9a–b).

Vertical wind shear was weak throughout the troposphere. We use the change in system-relative across-line wind speed ( $\Delta U$ ) from 0.5 to 2 km AGL as a measure of low-level wind shear,  $\Delta U$  from 0.5 to 5 km AGL as a measure of midlevel wind shear, and  $\Delta U$  from 0.5 to 10 km AGL as a measure of deep-layer wind shear. For our measures of environmental wind shear, we exclude wind values from the lowest 500-m layer, which are highly variable in time in our dataset, and are clearly influenced by surface-layer frictional effects. James et al. (2005) also cautioned against using near-surface winds for calculations of environmental shear. For S1, all measures of  $\Delta U$  were  $\sim 10 \text{ m s}^{-1}$  (Fig. 9d–f), which is similar to the mean 0–2 km wind vector difference in the MCS climatology of Cohen et al. (2007).

S2 (Fig. 8b) was released 36 minutes after S1. (In Fig. 8b, and subsequent skew-T diagrams, the new sounding is shown by thick black lines and the previous sounding in the series is shown by thin black lines.) By this time, an optically thick cirrus anvil had passed between the launch site and the sun. The near-surface temperature profile was now statically stable in the lowest 100 m, presumably because surface sensible heat flux reversed sign after solar heating was removed. S2 also showed evidence of deep-layer lifting; compared to S1, the cap at 775 mb was almost completely gone, and the mixing ratio was higher from the surface to 750 mb. A comparison of equivalent potential temperature ( $\theta_e$ ) profiles (Fig. 10) suggests that 200–300 m of lifting had occurred below 2.5 km AGL (assuming conservation of  $\theta_e$  during parcel displacements).

The cirrus anvil was apparent in S2 as a nearly saturated layer above 300 mb (Fig. 8b). CAPE in S2 was  $3890 \text{ J kg}^{-1}$  ( $\sim 380 \text{ J kg}^{-1}$  lower than S1, owing to warming in upper levels) and CIN was now only  $4 \text{ J kg}^{-1}$  (about half of S1, owing to cooling near 775 mb) (see also Figs. 9a–b). Measures of environmental wind shear had changed little (Fig. 9d–f).

In section 2, we noted a 0.9 mb increase in surface pressure in 15 minutes at the Cherokee mesonet station (Fig. 5b), exactly preceding the launch of S2. Using soundings S1 and S2 we can check whether this pressure change was attributable to cooling in low levels. Integrating the hydrostatic equation downward from 600 mb, we find the temperature and moisture

changes in S2 contributed to a 0.6 mb increase in surface pressure, which is consistent with the argument that most of the pressure increase is attributable to cooling in low-levels between the time of S1 and S2.

S3 (Fig. 8c) was released 27 minutes after S2. Compared to S2, S3 was more moist at almost every level above 800 mb. The cloud base of the cirrus anvil was lower, at  $\sim 350$  mb. Based on observations by the authors on this day, the S3 sonde likely passed through a mammatus layer at this cloud base. The nearly mixed subsaturated layer below an approximately moist-adiabatic saturated layer are common features of mammatus environments (e.g., Schultz and Trapp 2003; Schultz et al. 2006; Kanak et al. 2008).

Between 800 and 900 mb, the difference between S2 and S3 is perhaps surprising: this layer was substantially warmer and drier. The 1.2-K increase in temperature at 1.5-km AGL (see Fig. 11) was a departure from the cooling that was observed between S1 and S2. Comparing profiles of  $\theta_e$  (dashed and dotted lines in Fig. 10), and assuming  $\theta_e$  is conserved during parcel displacements, there is evidence of 200–300 m of subsidence in this layer.

This warming/subsidence feature might be attributable to a propagating gravity wave, although the origin of the gravity wave and the exact type of wave are difficult to discern from this dataset. The gravity wave could have been a type of low-frequency wave that is triggered by convective initiation (sometimes referred to as the “n=1” mode); see, for example, Nicholls et al. (1991), Mapes (1993), and Fovell (2002). This type of gravity wave is associated with subsidence that can cause warming of  $O(1\text{ K})$ . However, compared to these previous modeling studies, the warming in this case is shallow ( $\sim 1$  km deep) and is located at comparatively lower levels (that is, it’s not in the middle troposphere). These differences could be attributable to the presence of multiple dry-adiabatic layers (Fig. 8), within which gravity waves cannot propagate; in contrast, these earlier studies had continuous stably stratified environments over a deep layer. The warming in S3 might also be attributable to a high-frequency gravity wave that is trapped beneath the elevated mixed layer. The study by Fovell et al. (2006) focused on this type of wave, which occurred at a similar level in their

numerical simulations.

The drop in surface pressure just before the launch of S3 is also consistent with the passage of a gravity wave (Adams-Selin and Johnson 2009). A similar pressure drop occurred in Oklahoma mesonet sites to the east and southeast of Cherokee (not shown), although the amplitude of the pressure change was progressively smaller with distance from Cherokee. The timing of this feature in mesonet data was consistent with a wave moving southeastward at  $\sim 30 \text{ m s}^{-1}$ ; for reference, we note that wind speed in S2 was less than  $22 \text{ m s}^{-1}$  at all levels.

We conclude that the observed warming in S3 is reasonably consistent with the passage of a gravity wave, although this conclusion is admittedly difficult to support with the existing dataset. A numerical modeling study would probably be needed to evaluate more definitively the effects of gravity waves in this case.

Comparing now the  $\theta_e$  profiles at higher levels, there was evidence of  $\sim 500 \text{ m}$  of lifting (see dashed and dotted lines in Fig. 10). A midlevel cap near  $p = 500 \text{ mb}$  in S2 was at  $p = 450 \text{ mb}$  in S3 (Fig. 8c). Because of this *cooling* in mid levels, CAPE was  $3980 \text{ J kg}^{-1}$  ( $90 \text{ J kg}^{-1}$  higher than S2). Because of the abrupt *warming* at lower levels, CIN had increased to  $33 \text{ J kg}^{-1}$  (4 times higher than S2) (Figs. 9a–b). A backing of the winds at  $800 \text{ mb}$  had increased  $\Delta U$  in low levels (from  $9$  to  $11 \text{ m s}^{-1}$ ; see Fig. 9d). A slight backing of the wind along with a small wind-speed increase in the anvil cloud layer resulted in a sharp increase in deep-layer  $\Delta U$  (from  $12$  to  $23 \text{ m s}^{-1}$ ; see Fig. 9f).

S4 (Fig. 8d) and S5 (Fig. 12) were released concurrently, 26 minutes after S3. Both were released ahead of the squall line’s gust front: see Fig. 3a. Because of system propagation and advection of the sonde by winds, the mid and upper levels of these two soundings sampled the convective region of the squall line (see Fig. 4; sonde trajectories are located at  $x_c < 0$  in upper levels). We focus first on sounding data in low levels, which sampled the pre-squall-line environment.

The most notable difference between S3 and S4 is the humidity: S4 was more moist (10–40% in terms of relative humidity, 5–10 K in terms of dewpoint temperature) over a deep layer

from 800–600 mb. Precipitable water was 30% higher in S4 compared to S1 (Fig. 9c). There had also been substantial cooling (1–2 K) in the layer between 625 and 775 mb (Fig. 11). Both features are consistent with deep-layer lifting, perhaps by the approaching squall-line cold pool. Modeling studies have shown that low-frequency gravity waves (particularly the “n=2” mode) can be a cause of pre-line cooling and moistening (Fovell 2002; Fovell et al. 2006) and it is possible that such a mechanism contributed to the observed behavior. Profiles of  $\theta_e$  (Fig. 10) suggest that 500 m of lifting occurred between the 800 and 600 mb levels. This cooling led to an increase of CAPE to  $4235 \text{ J kg}^{-1}$  (a  $250 \text{ J kg}^{-1}$  increase from S3). Because conditions below 1.75 km AGL remained nearly unchanged, the CIN was still rather high ( $32 \text{ J kg}^{-1}$ ) (Figs. 9a–b). Further increases in wind speed at 800 mb increased  $\Delta U$  in low levels (Fig. 9d). The upper-level winds (above 300 mb) in S4 were notably stronger than in S1–S3, although deep-layer measures of  $\Delta U$  (e.g., Fig. 9f) decreased slightly due to backing of near-surface winds.

To summarize this section of environmental soundings, we note that the evolution of the near-squall-line environment was rather complex, but was consistent with previous studies. The observations herein are roughly consistent with numerical simulations showing how gravity waves can modify the low-level environment ahead of squall lines (e.g., Fovell 2002; Fovell et al. 2006), and how blocking of the sun by cirrus outflow can have a significant effect on conditions near the surface (e.g., Markowski and Harrington 2005). These processes were captured in unprecedented detail by these soundings from VORTEX2.

## b. Convective region

S5 was launched only  $\sim 4$  km in front of the squall line gust front (Fig. 3a). Analysis of the sonde’s GPS tracking suggests that the sonde rose normally (i.e., typical sonde ascent of  $\sim 4 \text{ m s}^{-1}$ ) until 900 mb, and thereafter experienced more rapid ascent, likely associated with the squall line’s cold pool; note that this corresponds with  $x_c = 0$  on Fig. 4. From roughly 800 to 625 mb, the sonde measured saturated conditions with occasionally decreasing values

of  $\theta_e$  (Figs. 12 and 13). From closer analysis, we find that S5 documented a moist absolutely unstable layer (MAUL; Bryan and Fritsch 2000), in which  $\theta_e$  decreased in a continuously saturated layer 1.2 km deep (from 760 to 660 mb). The strongly superadiabatic layers at 620 mb and 490 mb are probably erroneous and are likely attributable to the “wet bulb effect” when a radiosonde exits a cloud (e.g., Slonaker et al. 1996, p. 351).

Vertical air velocity ( $w$ ) from S5 is plotted in Fig. 14 (black line), where  $4.3 \text{ m s}^{-1}$  was subtracted from the sonde’s actual ascent rate (which gives approximately zero mean air velocity near the surface). Average  $w$  is  $6 \text{ m s}^{-1}$  in the MAUL layer, and  $w$  is not lower than  $5 \text{ m s}^{-1}$  in the entire layer; such values are consistent with a deep layer of ascent associated with the approaching cold pool, which forms the MAUL by the process described in Bryan and Fritsch (2000).

Above 625 mb, the sonde’s trajectory and thermodynamic data were chaotic, and at times the sonde descended briefly. The sonde was clearly in the turbulent convective region at this time. It is interesting to note that  $\theta_e$  during the turbulent time period is always *lower* than  $\theta_e$  at the more laminar lower levels (Fig. 13). This observation suggests that convective updrafts were entraining lower- $\theta_e$  air from midlevels and/or that the sonde occasionally left convective updrafts. However, thermodynamic data are questionable in this layer owing to probable wetting and/or icing of the sensor; the magnitude of pressure/temperature/humidity errors is probably impossible to predict under these conditions. Thus, we cannot conclude with any confidence whether the sounding data in the convective updrafts shows evidence of entrainment. The maximum value of  $w$  from S5 was  $25 \text{ m s}^{-1}$  (Fig. 14), although this value might not be representative of actual convective updrafts for this event owing to occasional loss of satellite tracking, and possible icing of the sonde.

The next sounding, S6 (Fig. 15a), is probably the most interesting of the entire dataset. It was released approximately 11 minutes *after* the gust front passed the primary sounding site, but several minutes *before* precipitation started at the surface. Direct measurements of temperature and moisture above the ground are very rare in such locations, with the

exception of some tower-based observations up to  $\sim 0.5$  km AGL (e.g., Charba 1974; Goff 1976). Surface temperature at S6 was 4.7 K lower than it was at S4 (which was launched from the same site 32 minutes earlier). In fact, temperatures from S6 were lower than temperatures from S4 from the surface to 3.6 km AGL (at 630 mb). The sonde was only 8 km behind the surface gust front when it was at the top of this cool pool. The radar analysis (Fig. 4b) shows that the sonde did not experience precipitation until the top of the cold pool, and even then the reflectivity was only 10 dBZ. It thus seems unlikely that there were any significant precipitation-induced instrument errors in the cold pool.

The wind profile of S6 had strong northwesterly (ground-relative) flow exceeding 50 kts ( $26 \text{ m s}^{-1}$ ) in the middle of the cold pool. This wind speed was greater than the system propagation speed at this time, and so the sonde was moving *away* from the approaching precipitation (see also Fig. 4b). The sonde rose at a nominal ascent rate of  $\sim 5 \text{ m s}^{-1}$  from the surface to 850 mb. Thereafter, its ascent rate gradually increased. Vertical air velocity from S6 is shown in Fig. 14 (gray line), where  $5.1 \text{ m s}^{-1}$  has been subtracted from the sonde's actual ascent rate (which makes  $w$  approximately zero near the surface).

Values of  $\theta_e$  in the cold pool were nearly constant (dotted line in Fig. 13). In fact,  $\theta_e$  in low-levels of S6 was the same value as  $\theta_e$  from midlevels in the pre-storm environment (cf. solid line in Fig. 13). Thus, it is likely that the source region of the cold pool air was midlevel environmental air, consistent with the conclusion reached by Newton (1950) and many studies since. Because the cold pool had similar  $\theta_e$  to that of midlevel environmental air, this air likely descended undiluted. However, values of water vapor mixing ratio (not shown) were 4–8  $\text{g kg}^{-1}$  higher than midlevel environmental air, confirming that a large amount of evaporation had occurred.

Above 650 mb, the thermodynamic and velocity data from S6 were chaotic. Values of  $\theta_e$  above 650 mb were roughly similar in S5 and S6 (Fig. 13), although thermodynamic data are questionable in this layer owing to probable wetting and/or icing of the sensor. Maximum  $w$  for S6 was  $20 \text{ m s}^{-1}$  (at 7.3 km AGL) (Fig 14).

To analyze further the properties of the squall line cold pool, we plot buoyancy ( $B$ ) in Fig. 16, where

$$B = g \left[ \frac{\theta - \bar{\theta}}{\bar{\theta}} + 0.608 (q_v - \bar{q}_v) \right], \quad (1)$$

$\theta$  is potential temperature,  $q_v$  is water vapor mixing ratio, and overbars denote reference (i.e., environmental) values. We neglect the contribution to  $B$  from condensate because it was likely negligible in the cold pool for S6. We calculate profiles of  $B$  from S6 using two different reference profiles: the far environment (S1, Fig. 16a) and the near environment (S4, Fig. 16b). Because the near environment had cooled substantially in low levels (as discussed in the previous section), the diagnosed cold pool depth (located where  $B = 0$ ) is only 3.6 km in Fig. 16b, as opposed to 4.0 km in Fig. 16a (Table 2).

A common measure of total cold pool intensity is the variable  $C$ , calculated by

$$C^2 = -2 \int_0^h B dz, \quad (2)$$

where  $h$  is the height at which  $B$  first equals zero.  $C^2$  is proportional to the surface hydrostatic pressure increase from a layer of relatively cool air. Although  $C$  is also, in principle, proportional to the maximum possible propagation speed of a gust front, we use  $C$  herein strictly as a convenient measure of vertically integrated buoyancy (i.e., total cold pool intensity). When using S1 as the reference we find  $C = 37.7 \text{ m s}^{-1}$ , and using S4 we find  $C = 34.4 \text{ m s}^{-1}$  (Table 3). These values are comparable to the largest values observed in MCSs during BAMEX (Bryan et al. 2004, 2005), but are notably greater than values reported in some comprehensive numerical modeling studies (e.g., Weisman and Rotunno 2004; Stensrud et al. 2005; Bryan et al. 2006). A cold pool of similar intensity was produced in a numerical simulation by James et al. (2005), but only when they used an environmental sounding having a relatively deep mixed layer. Of course, the analysis herein is from only one case; hence, it is unknown whether such large values of cold pool depth and intensity are extreme or common in the central United States. The issue is relevant, considering recent

debate about the intensity of cold pools and its relationship to MCS structure and intensity (e.g., Stensrud et al. 2005; Weisman and Rotunno 2005).

We now consider system-relative line-normal wind profiles ( $U$ ), which are shown in Fig. 17. The motion of the squall line is assumed to be southeastward at  $16 \text{ m s}^{-1}$ . As discussed earlier, environmental wind shear was weak throughout most of the troposphere, but not in low levels (see solid and dashed lines in Fig. 17a, and also Fig. 9d). Regardless of the depth over which  $\Delta U$  is measured, it is clear that  $C$  was significantly greater than  $\Delta U$  (by roughly a factor of 3;  $C$  was of order  $30 \text{ m s}^{-1}$ ,  $\Delta U$  was of order  $10 \text{ m s}^{-1}$ ). The theory for squall line structure and intensity proposed by Rotunno et al. (1988) and revised by Weisman and Rotunno (2004) (hereafter RKW Theory) predicts that squall lines should be tilted upshear when the cold pool intensity ( $C$ ) is much greater than the change in line-normal wind speed with height (i.e., the shear) in low- to mid-levels (specifically,  $\Delta U$  measured from near the surface to a height of order  $h$ ). Consistent with theory, the squall line’s convective region and updrafts were far behind the surface gust front (by  $\sim 20 \text{ km}$ ). This result is broadly consistent with a similar analysis by Bryan et al. (2004). We do not evaluate aspects of system intensity as it relates to RKW Theory, which would probably require multiple observations of  $C$  and  $\Delta U$  at various stages of this storm’s lifecycle, or perhaps measurements of  $C$  and  $\Delta U$  from other squall lines.

Notably, the pre-line environmental flow was *towards* the squall line (i.e.,  $U$  was negative) at all levels below 10 km AGL (solid and dashed lines in Fig. 17a). It is thus likely that some of the low- $\theta_e$  air in the cold pool came from *ahead* of the squall line. Specifically, midlevel air approached the squall line, likely passed between convective cells in the convective region [i.e., the “crossover zone” discussed by Zipser (1977)], and then descended into the cold pool due to evaporation and/or melting. Further details of this process were discussed by Zipser (1977). We also note that this process occurred in the numerical simulations of upshear-tilted squall lines by Rotunno et al. (1988).

In the cold pool (i.e., in S6),  $U$  was positive from 0.1 to 2.9 km AGL (dotted line in

Fig. 17a). Above 2.3 km AGL, flow in the cold pool was negative. (Recall that the cold pool was 3.6–4.0 km deep; Fig. 16). Hence, there was an overturning circulation in the cold pool, in which near-surface air moved toward the front of the squall line, but moved away from the squall line in the upper-half of the cold pool.

Of particular interest here is the change in  $U$  over time ( $\delta U$ ) as the cold pool passed, which is shown in Fig. 17b. The sounding data had a maximum value of  $\delta U$  of  $\sim 28 \text{ m s}^{-1}$  at 0.3 km AGL; closer to the surface,  $\delta U$  was slightly lower because of surface friction. Theoretically,  $C$  is proportional to  $\delta U$  at the surface; but in this case,  $\delta U$  ( $\approx 28 \text{ m s}^{-1}$ ) was clearly smaller than  $C$  ( $\approx 37 \text{ m s}^{-1}$ ). Technically,  $\delta U$  should be somewhat smaller than  $C$  because of the “finite channel depth effects” identified by Benjamin (1968) [see also Klemp et al. (1994)], but which are not considered herein. However, the difference might also be attributable to the neglect of the warm anomaly aloft. That is, the warm anomaly aloft lowers surface pressure hydrostatically, but the warm anomaly is not accounted for in (2) because integration stops at the top of the cold pool. If we perform the integral in (2) to the top of available sounding data, we find  $C \approx 27 \text{ m s}^{-1}$ , which is very close to  $\delta U$ ; this result suggests that the warm anomaly aloft influences near-surface winds and, likely, propagation speeds of convective systems. Additional discussion of methods to calculate and interpret  $C$  (including the need to sometimes integrate buoyancy over a deep layer) can be found in Trier et al. (2006).

To conclude this analysis of the convective region, we note that the cold pool depth ( $h$ ) was 3.6–4.0 km, depending on which sounding was used to define the environment, and the cold pool intensity (as measured by  $C$ ) was 34–38  $\text{m s}^{-1}$ . Both values ( $h$  and  $C$ ) are unusually large compared to previous idealized numerical simulations and compared to tropical oceanic squall lines, although these values are consistent with several convective systems observed during BAMEX (Bryan et al. 2004, 2005; Ahijevych et al. 2006). We also evaluated the component of RKW Theory that addresses squall line structure. The system was tilted upshear, and RKW Theory predicts  $C > \Delta U$  for this type of squall line; consistent with

theory, we found measured values of  $C$  were about three times larger than  $\Delta U$  in low- to midlevels.

### c. Stratiform region

After the launch of S6, rawinsonde launches were suspended temporarily owing to safety concerns. Rain, wind gusts, and lightning lasted for approximately 15 minutes at the primary sounding site. After these conditions passed, sounding S7 was launched (37 minutes after S6). A radar scan from this time (Fig. 3c) showed that the launch site was in the trailing stratiform region of the squall line, just behind the convective region. S7 (Fig. 15b) showed a classic “onion” shape with a nearly saturated and approximately moist-adiabatic layer above the melting level overlying a sub-saturated and nearly mixed layer below the melting level. A layer of cool and nearly saturated air near the surface was very shallow ( $\sim 20$  m). Relative humidity at the Cherokee mesonet site was 85% at launch time, consistent with S7, but at 200 m AGL the relative humidity was only 42% in S7.

The depth of the cold pool in S7 was 4.7 km, using either S1 or S4 as the reference (Table 2). Hence, the cold pool was deeper farther toward the rear of the system. Interestingly, the top of the cold pool was  $\sim 1$  km *above* the melting level. Further details of the cold pool depth, and processes responsible for these values, are discussed in the next section. Measured values of  $C$  from S7 were  $35.5 \text{ m s}^{-1}$  (using S1 as the reference) and  $31.7 \text{ m s}^{-1}$  (using S4 as the reference) (Table 3); both values were slightly lower than  $C$  measured using S6. Wetting of the sensor may have caused temperatures to be erroneously low in the subsaturated layer below the melting level in S7–S9. The magnitude of the error is difficult to generalize because it depends on relative humidity and liquid-water content. It is possible that the depth and intensity of the cold pool in the trailing stratiform region is slightly overestimated by these rawinsonde observations.

An elevated jet of maximized windspeed, centered near 800 mb in S6, was still present in S7 at the same level as S6, although the ground-relative windspeed was slightly less ( $\sim 40$  kts

in S7 compared to  $\sim 50$  kts in S6). The near-surface wind was weak in S7, consistent with the Cherokee mesonet data (Fig. 5d). Considering the different winds, relative humidity, and  $\theta_e$  near the surface, this near-surface air seems to be from a different airstream than the dry and nearly-mixed air located a few hundred meters above the ground.

Soundings were then released every 15 minutes until sunset. (VORTEX2 operations ended at sunset owing to safety concerns.) S8 (Fig. 15c) was released 16 minutes after S7 and it showed many similar features. The elevated jet maximum was slightly weaker (by  $\sim 5$  kts). The bottom of the moist-adiabatic layer was slightly lower, now almost exactly at the melting layer. The most significantly different feature in S8 was greater moisture in the layer just above the surface: relative humidity was 10–15% higher in the 950–900 mb layer. The diagnosed cold pool depth for S8 was the same as S7 (Table 2), but  $C$  was slightly lower (by  $\sim 5\%$ ; Table 3).

S9 (Fig. 15d), the final VORTEX2 sounding from this event, was released 13 minutes after S8. The sonde only ascended to 550 mb and then began descending slowly (because, we presume, the rawinsonde became covered with ice). S9 was generally similar to S8. The relatively moist layer just above the surface had moistened further: relative humidity was 10–20% higher in the 950–800 mb layer than it was in S8.

## 4 Mesoscale analysis

To better visualize the entire rawinsonde dataset, we created a mesoscale analysis using the Barnes (1973) scheme. Based on average spacing of the sondes, and the high vertical resolution of the soundings, our analysis grid had 10 km horizontal grid spacing and 100 m vertical grid spacing. We used a two-pass Barnes scheme with a two-dimensional weighting function. The non-dimensional weighting parameter was 0.3. At grid points where observations were comparatively far away ( $> 20$  km), the analysis grid was left undefined.

The following analyses assume implicitly that the squall line and its near environment

were steady during this time period, which is a simplification of the actual events. The analyses also implicitly ignore along-line variability, and thus neglect the possible effects of advection along the line. These assumptions are required herein, given the dataset available.

An analysis of equivalent potential temperature ( $\theta_e$ ) is shown in Fig. 18a. This analysis shows a clear upshear tilt to the squall line convective region. In mid levels (6–8 km AGL), maximum  $\theta_e$  is located  $\sim 30$  km behind the surface gust front. Low-level ascent in the environment is revealed in this analysis by a slight tilt of the contours (for  $x_c > 0$  and  $z < 4$  km). In the trailing stratiform region,  $\theta_e$  was nearly well-mixed from near the surface to 12 km AGL, although  $\theta_e$  increased slightly with height. Near the surface in the stratiform region, the relatively moist air (documented earlier) appears in the analysis as a plume of high- $\theta_e$  air for  $x_c < -50$  km and  $z < 1.5$  km.

An analysis of relative humidity with respect to liquid is shown in Fig. 18b. Low-level ascent in the environment is reflected here by gradual increases in relative humidity, and a slight tilt of contours, from right-to-left. The warming near the top of the boundary layer, which we hypothesize to be related to a propagating gravity wave, can be seen here as a region of relatively lower relative humidity at  $x_c \approx 40$  km and  $z = 1.2$  km. Air in the squall-line’s cold pool ( $x_c < 0$ ) had generally low relative humidity, especially below 3 km AGL.

An analysis of system-relative cross-line wind speed is shown in Fig. 18c. For this analysis, the movement of the convective region (southeast at  $16 \text{ m s}^{-1}$ ) was used to define system-relative flow. Regions with positive cross-line winds are shown by gray shading. All salient features discussed earlier are captured by the analysis, including: negative system-relative flow throughout most of the environment; weak environmental wind shear in most of the troposphere, except below 2 km AGL; positive line-normal flow above 10 km AGL in the environment associated with the squall line’s upper-level outflow (i.e., the spreading cirrus cloud); the rear-inflow jet (centered near 2 km AGL for  $x_c < 0$ ); and accelerated flow towards the rear of the line (i.e., front-to-rear flow) in the convective region. This analysis suggests

an abrupt increase in the depth of the rear-inflow jet for  $x_c > -20$  km, although this feature was captured by only one sounding (S6). Analysis of Doppler radar data collected during VORTEX2 could be used to better determine the spatial extent of this rear-inflow jet. In the trailing stratiform region (for  $x_c < -40$  km), the elevated rear-inflow jet was only  $\sim 1$  km deep (from roughly 1 to 2 km AGL). However, we note that the back edge of the trailing stratiform region was moving slower than the convective region (by  $\sim 5$  m s $^{-1}$ ). In a similar analysis, but with flow relative to the back edge of the stratiform region (not shown), the rear-inflow jet was much deeper (from roughly 0.5 to 3 km AGL).

Analyses of buoyancy ( $B$ ) are shown in Fig. 19. For Fig. 19a we used the vertical profiles at  $x_c = 90$  km to define the environment, and for Fig. 19b we used the analysis at  $x_c = 10$  km. In Fig. 19a, the cooling and warming of environmental layers that were discussed earlier are clearly evident. In Fig. 19b, the edge of the squall line cold pool is shown more clearly. We note that both analyses show how the cold pool was *below* the melting level in the convective region ( $x_c > -30$  km), suggesting the probable influence of cooling from melting of snow/ice; but in the stratiform region the cold pool extended *above* the melting level, suggesting that sublimation was important for cooling the air in the 3.5–4.5 km level. Modeling studies by Stensrud et al. (1991) and Gallus and Johnson (1995) found sublimation to be small, but sometimes non-negligible, in the trailing stratiform region of squall lines.

An analysis of perturbation pressure is shown in Fig. 18d, where the profile at  $x_c = 90$  km was used as the reference. Consistent with surface observations from the Cherokee mesonet site, the analysis in Fig. 18d had a +5.5 mb pressure perturbation at the surface in the cold pool. Perturbation pressure was lowest at the top of the cold pool, near the melting level; the minimum analyzed value was  $-2.8$  mb. Relatively low pressure has long been recognized in observations of convective systems (e.g., LeMone 1983), although it is typically located at lower altitudes in tropical convective systems. This midlevel mesoscale low, and its position relative to the cross-line flow (Fig. 18c), is consistent with similar features shown in numerical simulations of bow echoes and squall-lines having trailing stratiform regions (e.g., Weisman

1993; Parker and Johnson 2004). Pre-line changes in near-surface pressure (of order 1 mb) are shown here to be shallow: they extended vertically only a few hundred m.

## 5 Discussion

Of the many questions raised by the analyses herein, perhaps the most pressing is: how representative are the analyses of cold pool depth and intensity? That is, how often are cold pools this deep ( $h \approx 4$  km) and this intense ( $C \approx 35$  m s<sup>-1</sup>)? The 4-km depth is consistent with the maximum theoretically possible cold pool depth (Bryan and Rotunno 2008), although this theory does not account for environmental wind shear or stably stratified environments. To our knowledge,  $C$  of order 35 m s<sup>-1</sup> has only been documented in a few numerical simulations (e.g., James et al. 2005; James and Markowski 2009), and only for CAPE  $\approx 4000$  J kg<sup>-1</sup>. On the other hand,  $C = 33$  m s<sup>-1</sup> was observed for the 10 June 2003 bow echo during BAMEX in an environment with CAPE  $\approx 2500$  J kg<sup>-1</sup> (Bryan et al. 2005).

Perhaps a cold pool “audit” would be beneficial to the community. Indeed, some recent studies (e.g., Stensrud et al. 2005; Engerer et al. 2008) have raised questions about the typical properties of midlatitude MCS cold pools, which are theoretically important for severe weather production, MCS propagation speeds, and MCS structure. In their study of mature convective systems in Oklahoma, Engerer et al. (2008) found average pressure rises of 4.5 mb and a maximum pressure rise of 9.4 mb; the  $\sim 5$  mb pressure rise for this case is near their average value, but it is unknown whether the depth or integrated buoyancy values found herein are also near average. Perhaps a future field project could be designed to conduct this audit. It might also be useful to capture observations at different stages in the lifecycle of convective systems. The soundings herein were collected early in the mature stage of the squall line. Observations at earlier and later stages might reveal different cold pool depth and intensity.

It would also be interesting to see if these analyses could be reproduced by numerical

simulations. The analyses herein raise several questions that are not easily addressed with the available observations, but could be investigated by numerical simulation. For example, what set of conditions are needed to produce cold pools of order 4 km deep? Are melting and sublimation important? How is a nearly uniform  $\theta_e$  profile obtained in the trailing stratiform region, and what model settings (e.g., physical parameterizations, resolution) are needed to reproduce this feature? The findings from such a study might help improve cloud-scale NWP forecasts.

The evolution of the pre-storm environment documented herein supports several recent theoretical and modeling studies about the impacts of anvil shading (e.g., Markowski and Harrington 2005) and the probable influence of convectively triggered gravity waves (e.g. Fovell 2002). The analyses also help address the question: what is an appropriate “proximity” sounding to define an MCS environment? In this case, CAPE is modified slightly (less than 10%) by the approaching squall line, but CIN increases markedly (by a factor of 4) within  $\sim 50$  km of the squall line. Both low-level (0.5–2 km) and deep-layer (0.5–10 km) measures of vertical wind shear increased markedly (by a factor of 2) within  $\sim 100$  km of the squall line (although for different reasons). Consequently, the “base state” soundings used to initialize idealized model simulations are probably not representative of the environment measured near convective systems (i.e., within  $\sim 100$  km) because they have not yet been modified by gravity waves and upper-level cirrus anvils.

Finally, one might ask: why aren’t analyses such as these more common? The answer is related to our data collection strategy for this event. First, we used mobile sounding platforms. Most historical studies used fixed sounding sites, such as the NSSL mesonetwork of the 1960s and 70s, and the PRE-STORM array in 1985; these studies depended on a convective system happening in one location, which reduces the number of possible cases. In contrast, we were able to position our observing system in an optimum location to collect these observations. Second, we had four sounding systems available to us, and so we were able to launch and track multiple sondes at the same time. This allowed us to decrease our

launch interval to 15–30 minutes, as opposed to the  $\sim 90$  minute launch interval for mobile sounding launches during BAMEX. (It takes about 90 minutes to prepare, launch, and track a sounding to the upper troposphere.) Third, the use of ground-based sounding systems allowed us to obtain sounding data very close to the convective region. It is too hazardous to obtain direct thermodynamic measurements from aircraft in high-CAPE environments. In contrast, a wealth of *in situ* aircraft observations have been obtained and analyzed for tropical environments; but it has been rare to collect direct thermodynamic measurements with high spatial resolution in mid-latitude continental convective systems.

One shortcoming of our analyses is the lack of soundings in the middle of the convective region. This shortcoming was attributable to safety concerns regarding the preparation and launch of sondes during lightning and heavy rain, because some of our equipment had to be operated outside vehicles. This problem could be addressed in the future by designing more equipment to be used inside the vehicle, or perhaps by using a portable shelter that could be extended from parked vehicles.

## 6 Summary

In this article, we analyze a novel set of rawinsonde observations collected before and during passage of a squall line in Oklahoma during VORTEX2. Nine soundings were released within a 3-hour time period. They documented, with unprecedented resolution, the changes in the pre-storm environment ahead of the squall line, and the internal thermodynamic and kinematic structure of the squall line.

In the pre-storm environment, the observations documented low-level cooling associated with the sudden decrease in solar insolation by a thick cirrus anvil. The cooling was  $\sim 500$  m deep, and the surface pressure increased by 1 mb in  $\sim 15$  min. Later, rapid warming and drying just above the top of the boundary layer were documented by the sounding data. Based on idealized modeling studies, and surface observations from the Oklahoma Mesonet,

it seems likely that a gravity wave passed by the sounding site. A 1-mb decrease in surface pressure, and a marked increase in convective inhibition, also happened at this time. Numerical modeling would be needed to better evaluate whether a convectively triggered gravity wave could have caused these changes in this environment. Later, as the squall line approached the sounding site, measures of low-level and deep-layer vertical wind shear increased. All of these features have been documented by past studies using theory and numerical simulations, but the observations herein might be unique.

A sounding released  $\sim 4$  km ahead of the squall line gust front documented a moist absolutely unstable layer (MAUL) in a rapidly ascending environment (vertical air velocity  $\sim 6 \text{ m s}^{-1}$  over a 2-km deep layer). Evidence suggests that weak lifting of the environment extended  $\sim 75$  km ahead of the squall line.

One sounding was released after the gust front passed but before precipitation was observed at the surface. This sounding documented a 3.6–4.0 km deep cold pool (depending on which environmental sounding is used as a reference). A direct measure of cold pool intensity ( $C$ ) from the soundings was compared to a measure of environmental vertical wind shear ( $\Delta U$ ); the analysis supports the component of “RKW Theory” that addresses system structure because the squall line was tilted upshear, consistent with  $C > \Delta U$  (Rotunno et al. 1988; Weisman and Rotunno 2004).

In the trailing stratiform region, equivalent potential temperature was almost constant from near the surface to near the tropopause, consistent with previous studies. The cold pool was deeper in the stratiform region (than in the convective region), and the top of the cold pool was above the melting layer, which suggests that sublimation was important for this case. An elevated rear-inflow jet was located just below the melting level.

The analyses herein should be valuable for assessing numerical model simulations and analytic theories. The analyses show that changes in the pre-storm environment (e.g., by gravity waves) are consistent with past modeling studies, but that some squall-line cold pools in the central United States are deeper and stronger than cold pools in some recent idealized

modeling studies. However, only one squall line was observed during the 2009 phase of VORTEX2, and so we cannot say whether these findings are common to midlatitude squall lines, or whether the results are rare. A future field project should be conducted to obtain a better sense of the climatology of midlatitude MCS cold pool properties.

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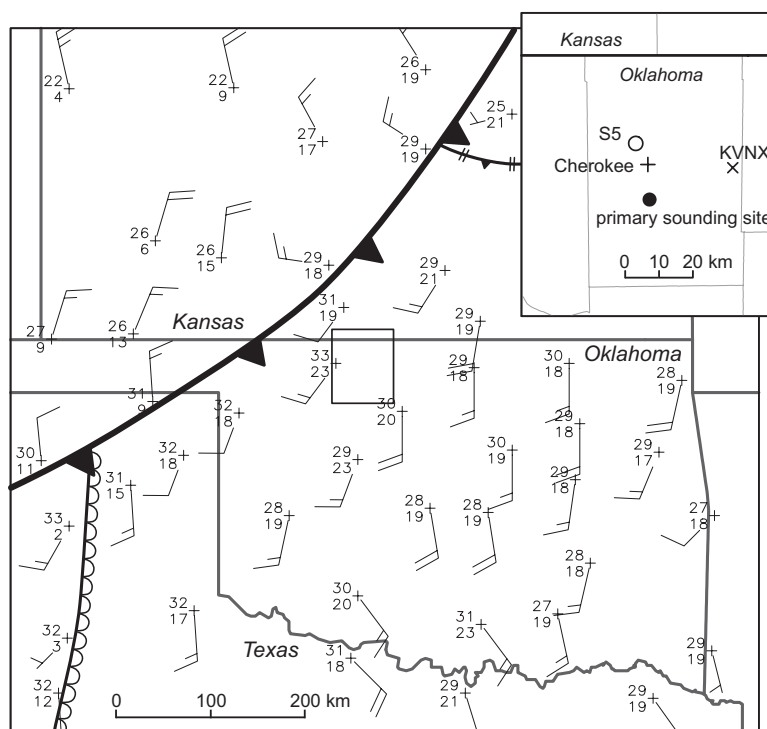


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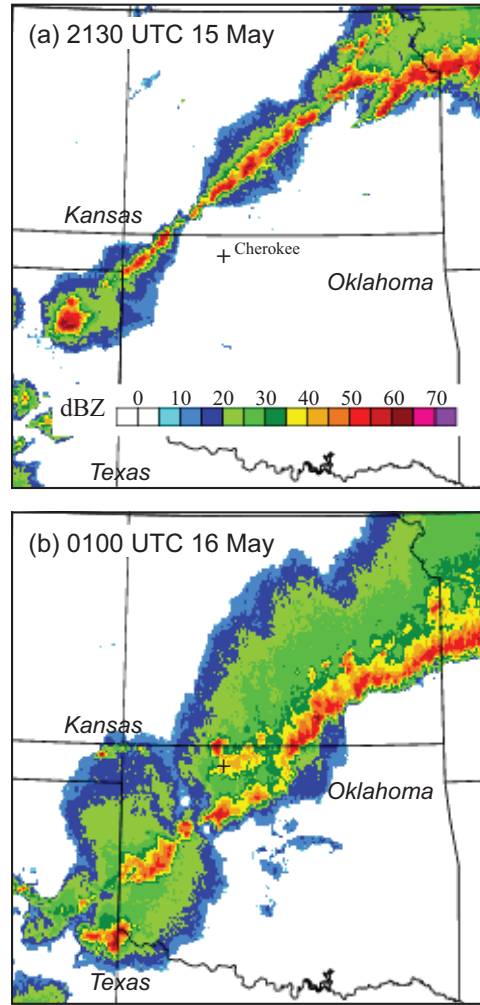


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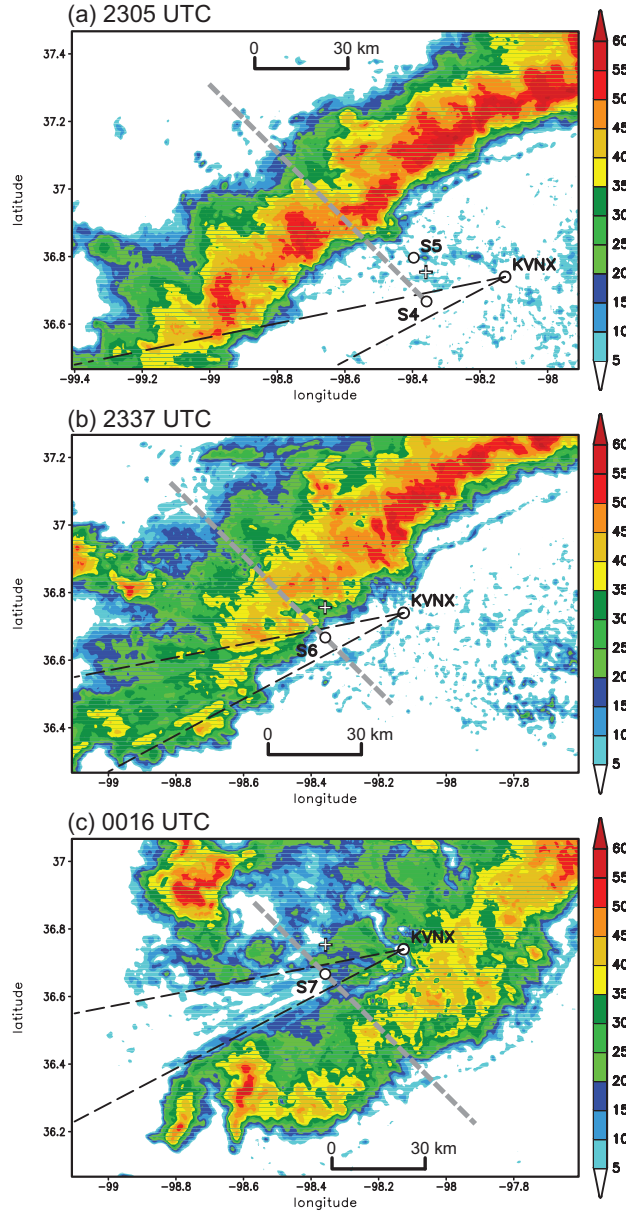


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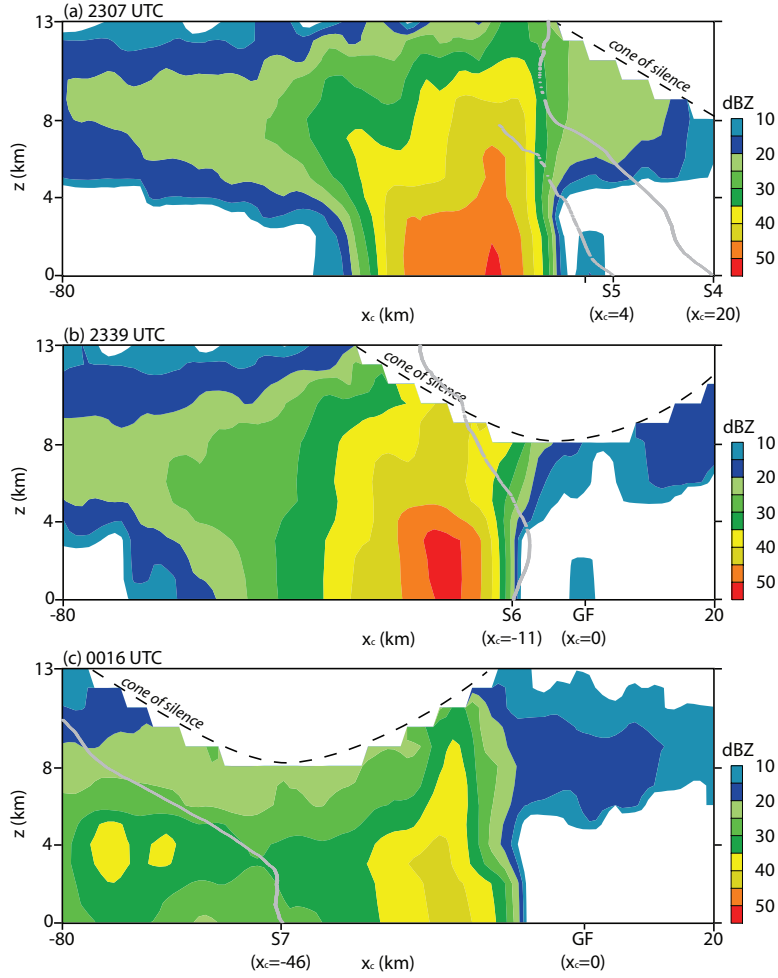


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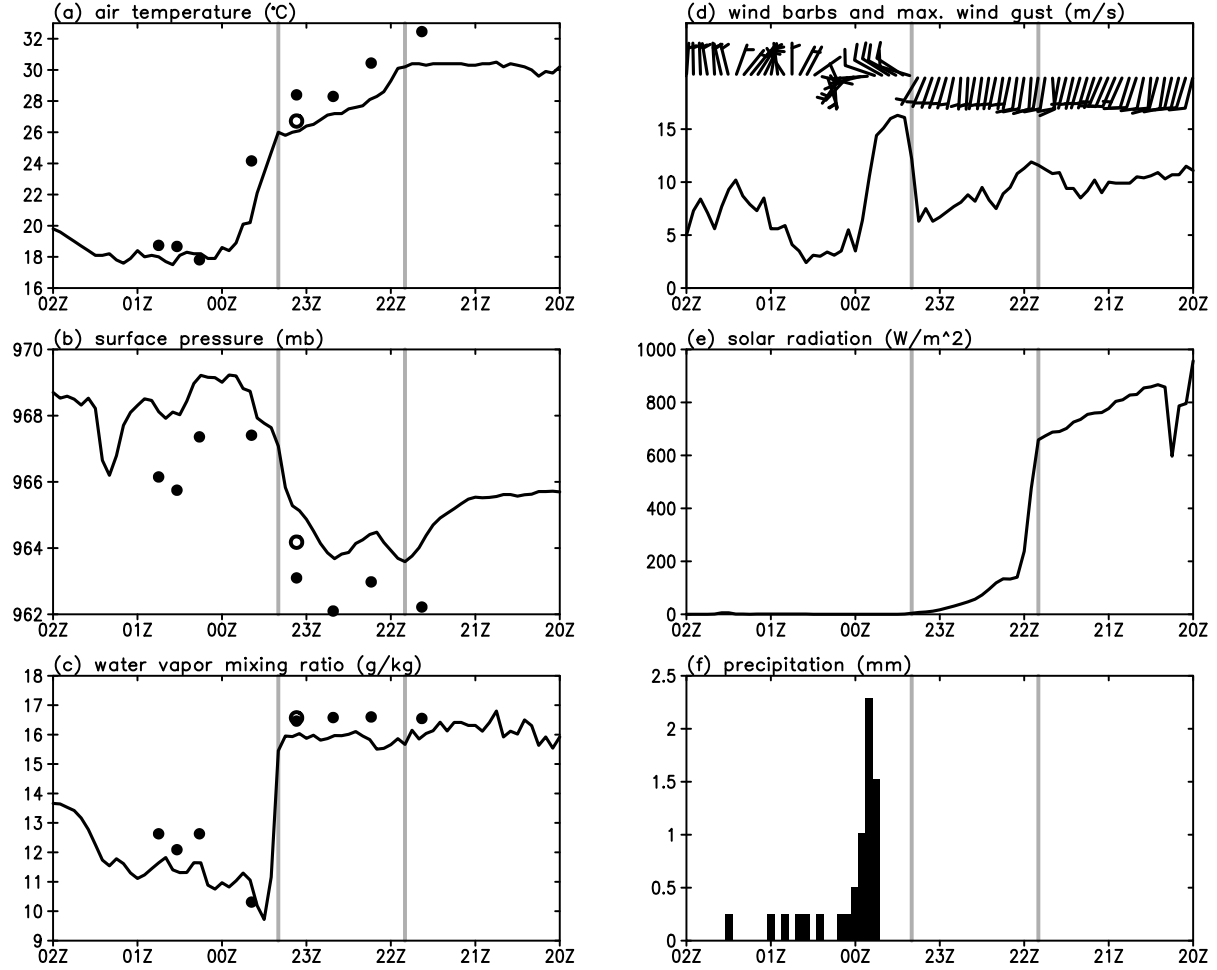


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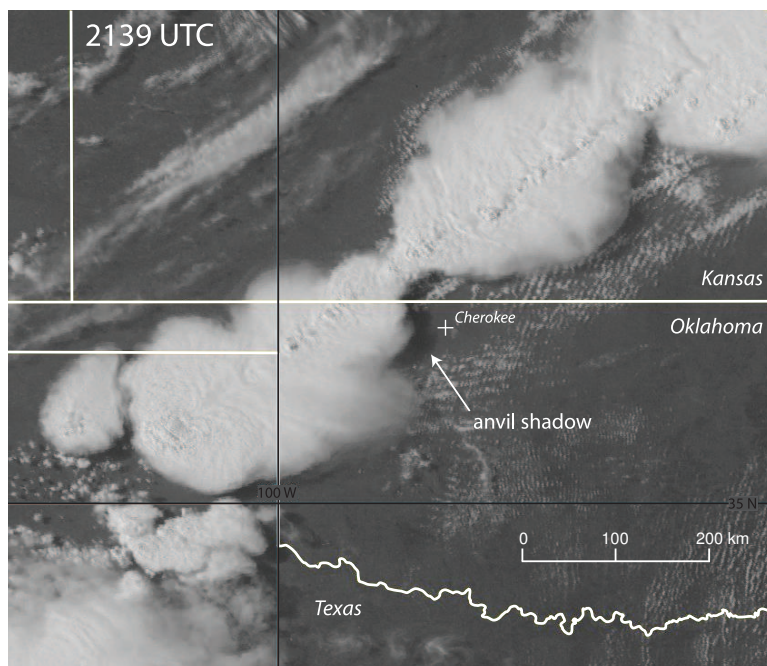


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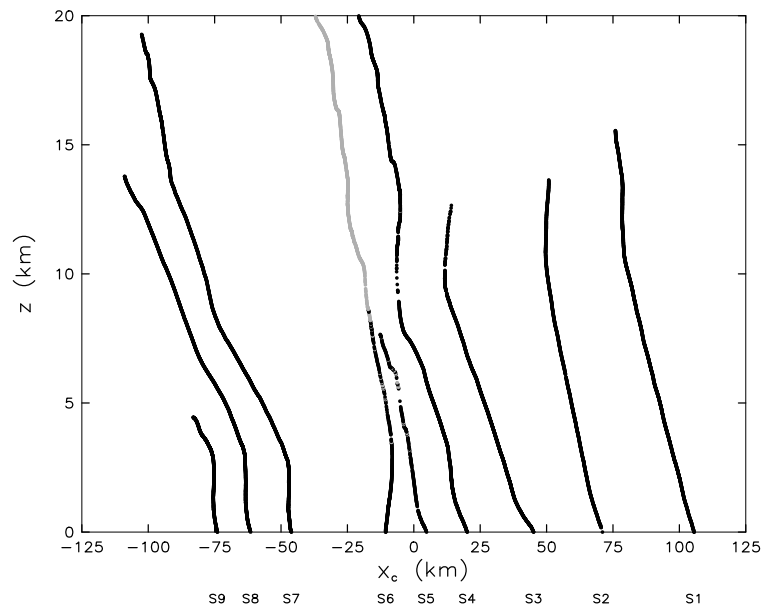


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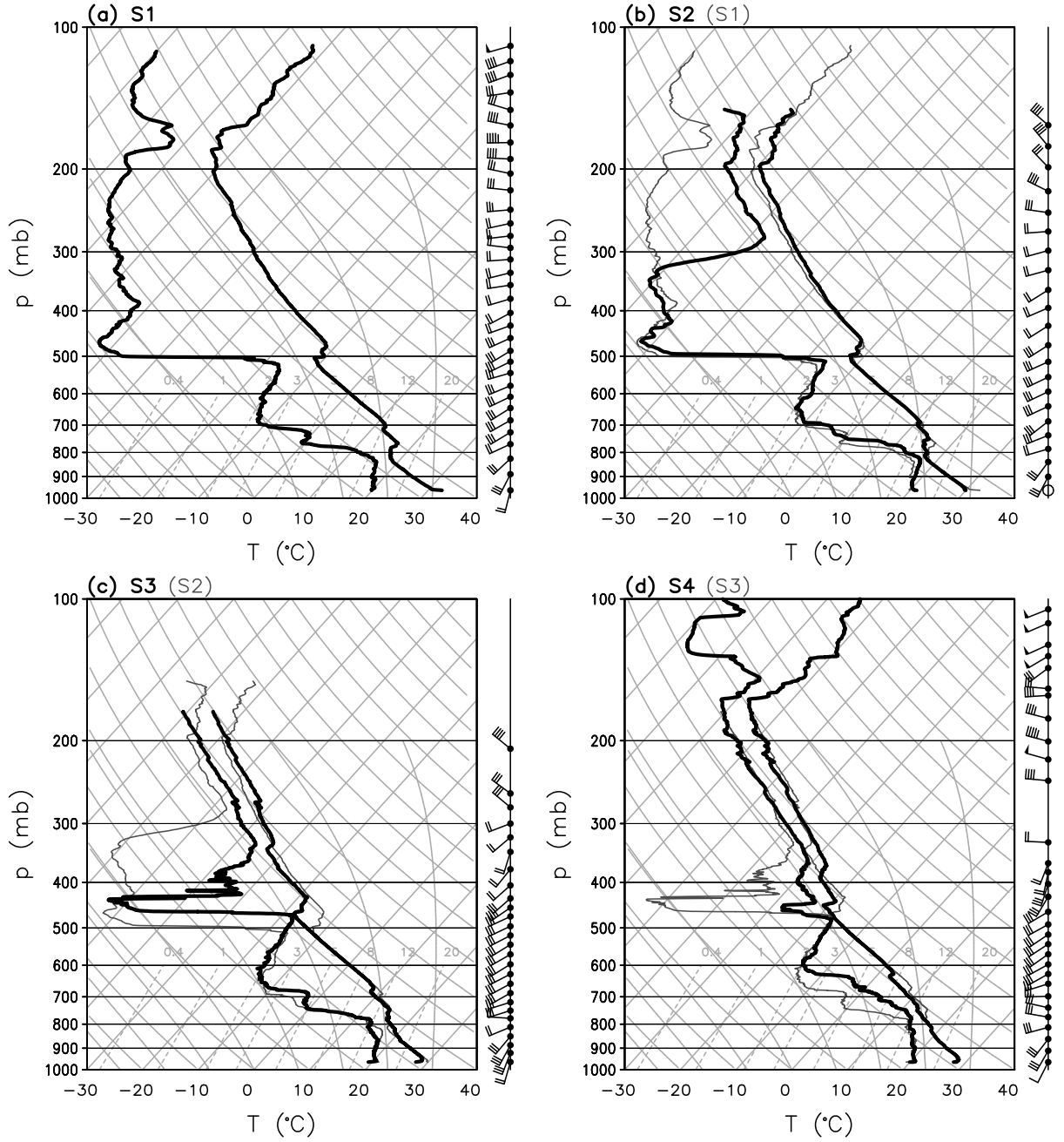


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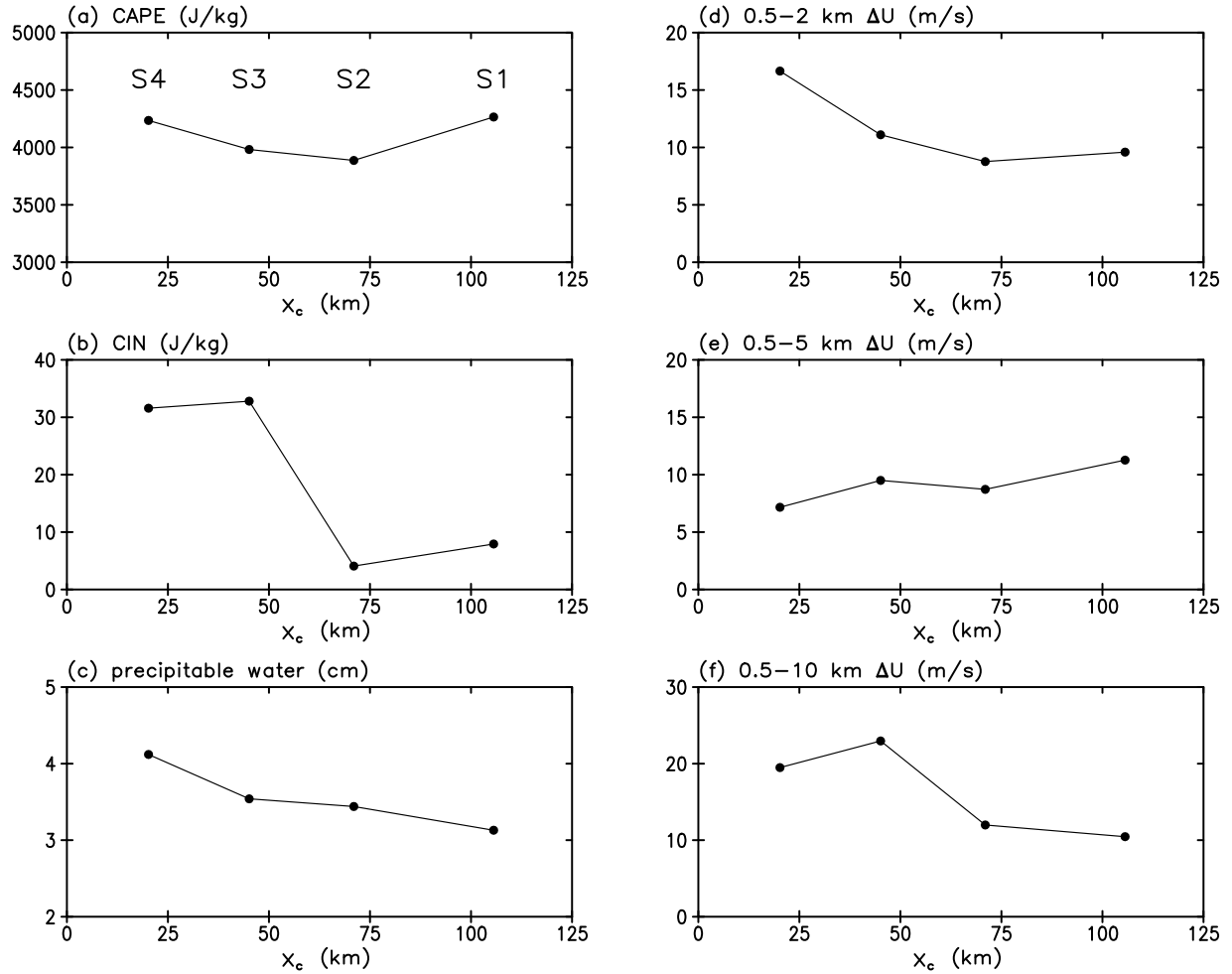


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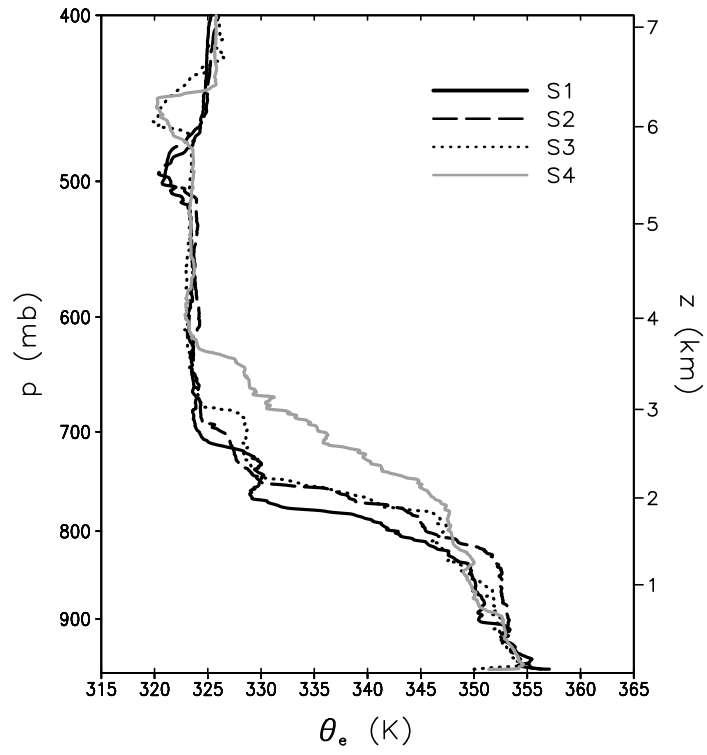


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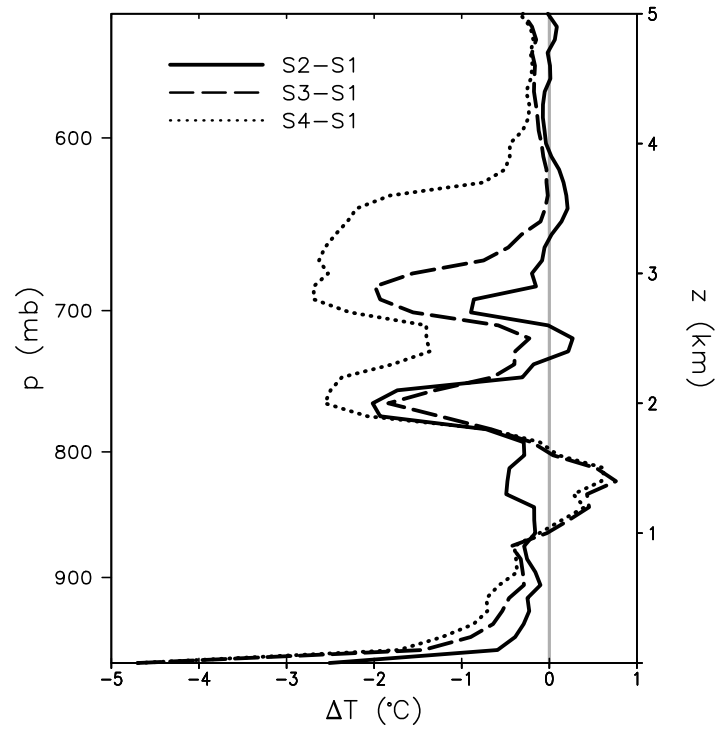


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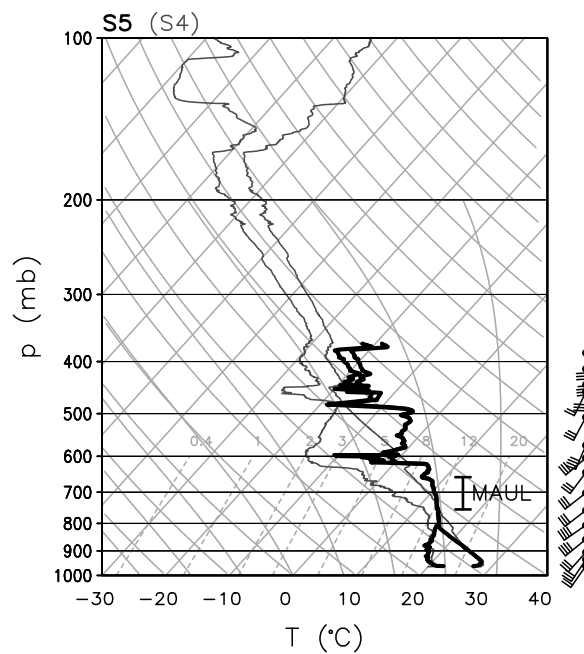


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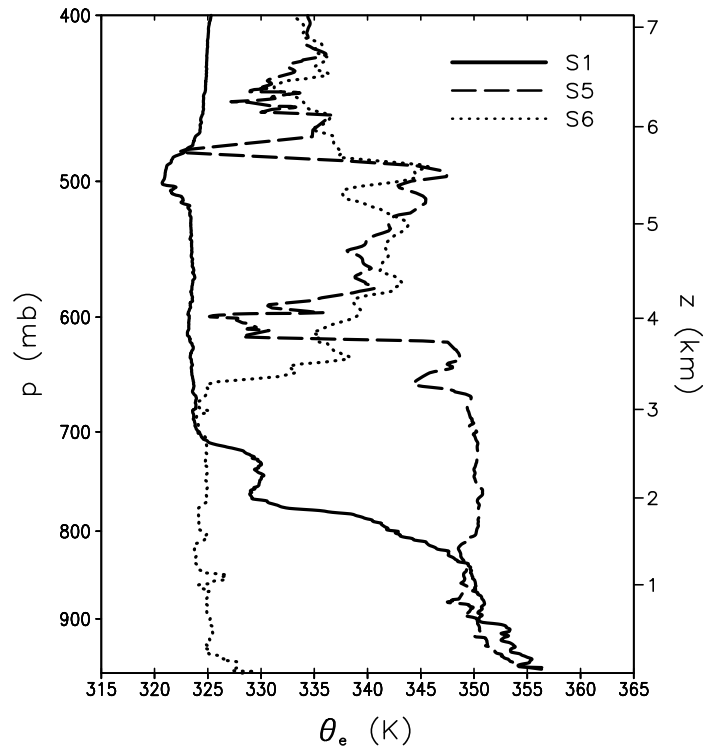


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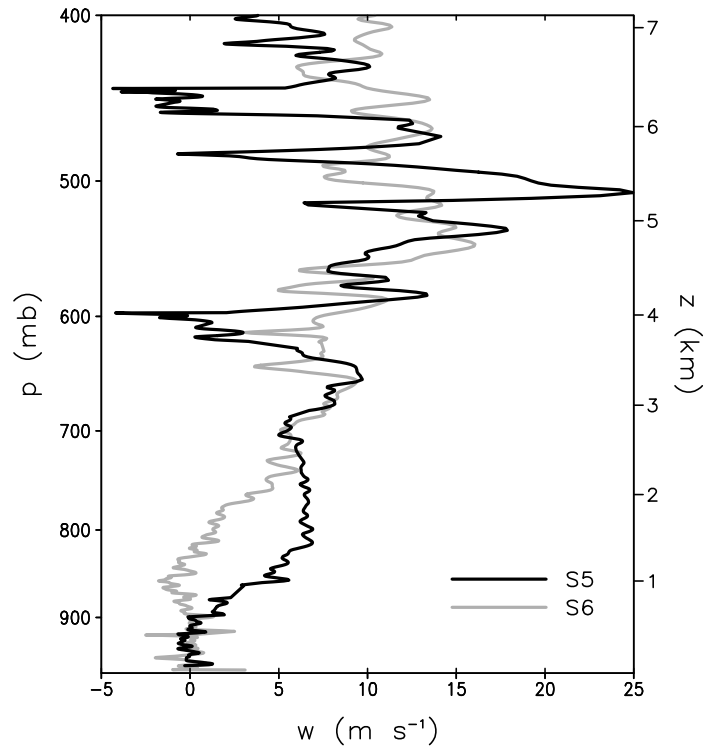


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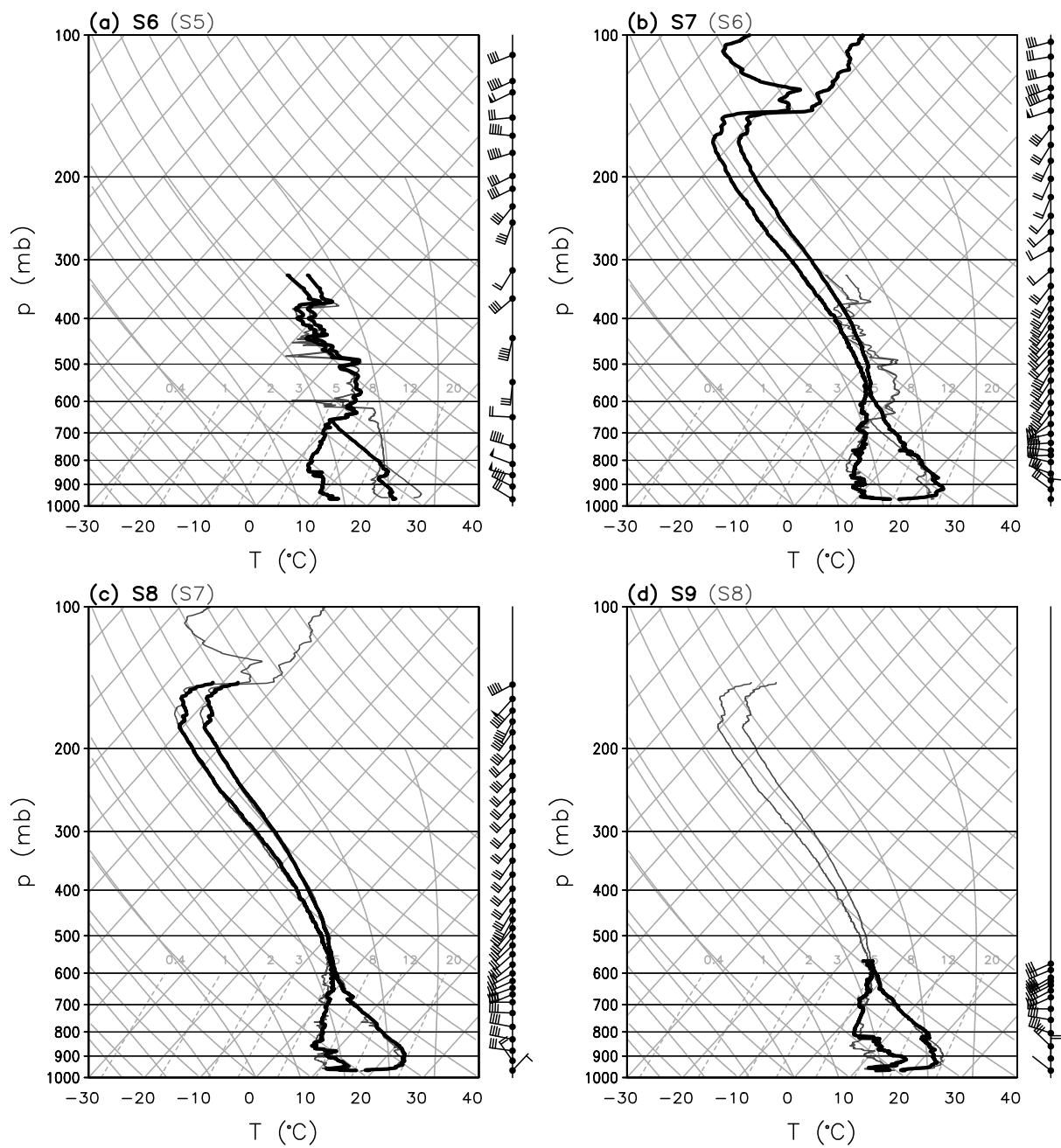


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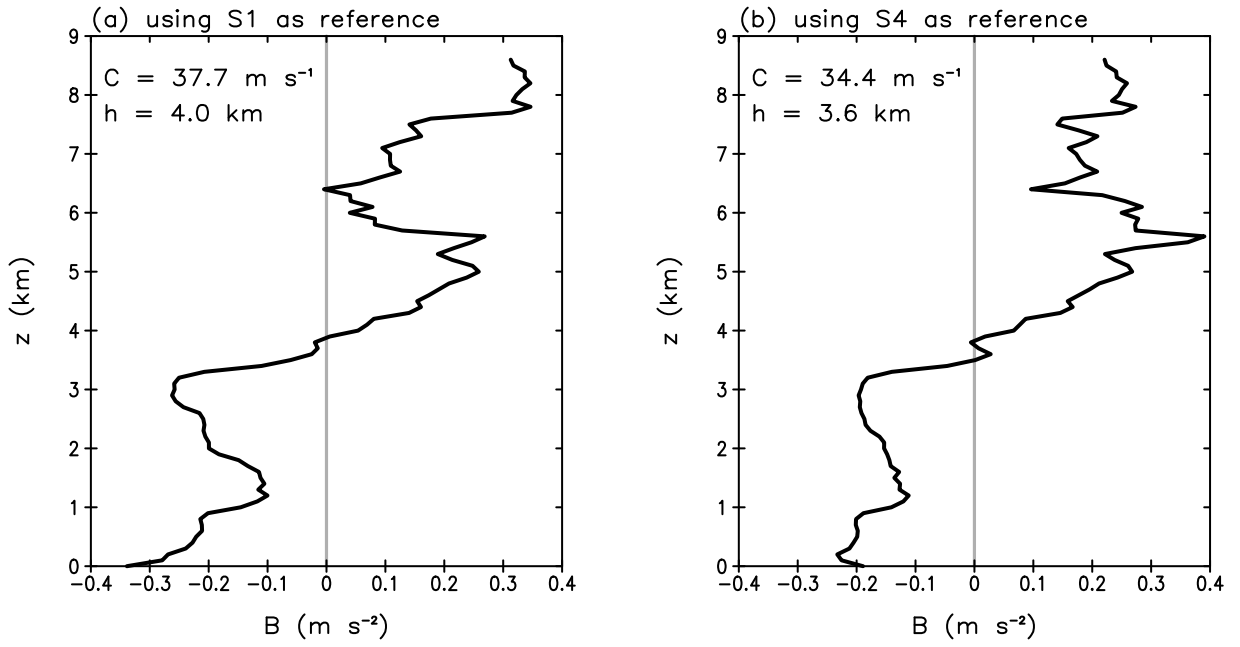


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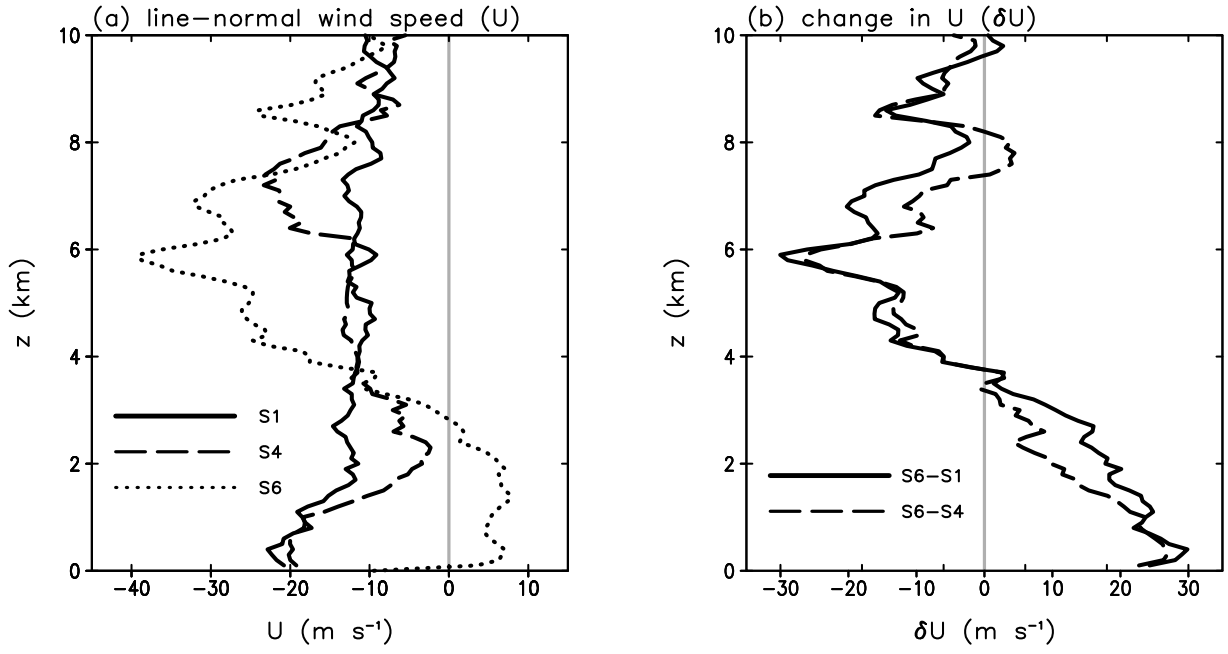


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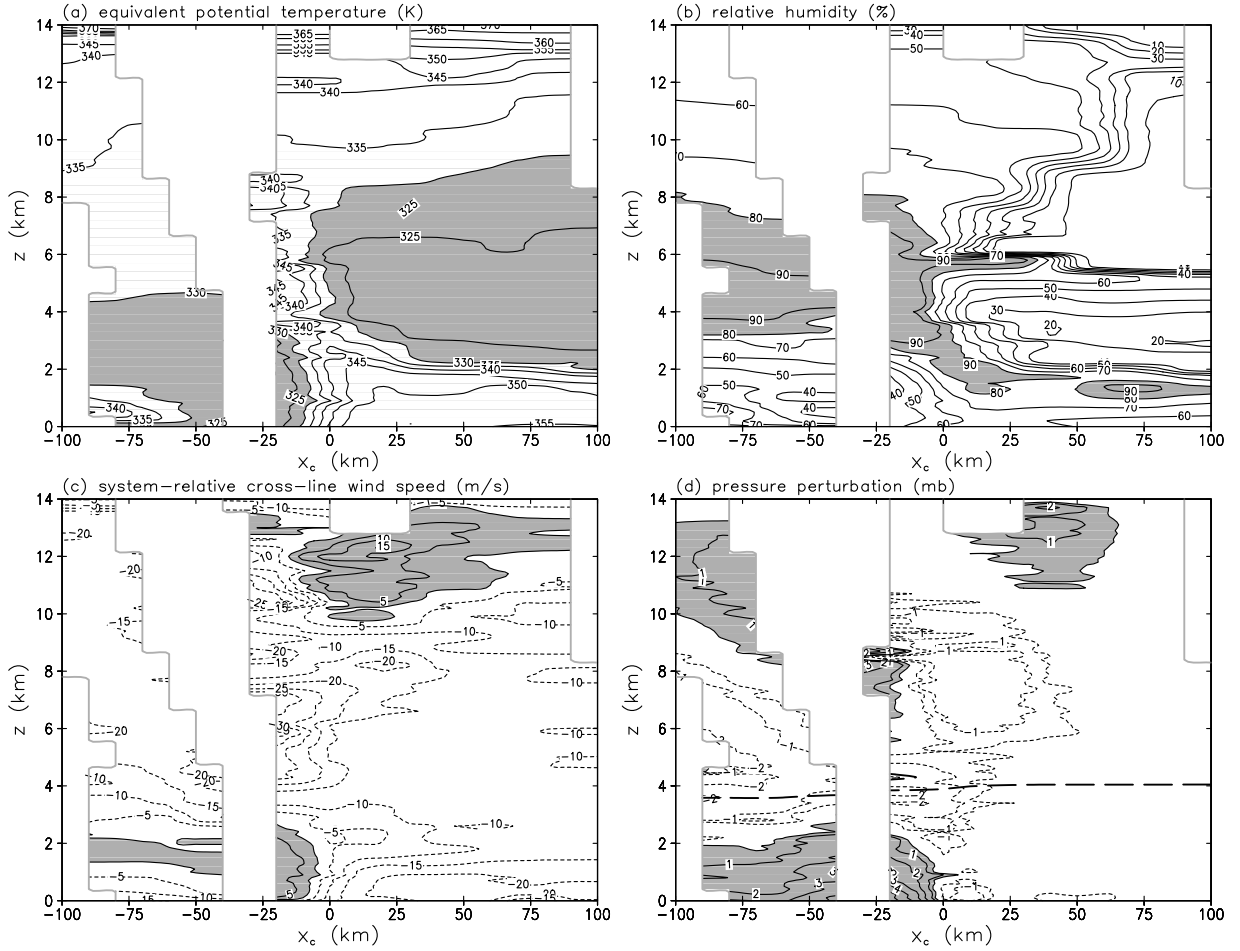


FIG. 18: Objective analyses as a function of across-line distance from the surface gust front ( $x_c$ ) and height AGL ( $z$ ): (a) equivalent potential temperature (contour interval is 5 K, and values less than 330 K are shaded); (b) relative humidity with respect to liquid (contour interval is 10%, and values greater than 80% are shaded); (c) system-relative cross-line wind speed (contour interval is 5 m s<sup>-1</sup>, negative contours are dashed, and values greater than zero are shaded); (d) pressure perturbation, relative to profile at  $x_c = +90$  km (contour interval is 1 mb except intermediate contours of  $\pm 0.5$  mb are included, the zero contour is excluded, and values greater than +0.5 mb are shaded). In panel (d), the melting level ( $T = 0$  °C) is shown as a thick dashed line. In all panels, areas with insufficient data are left blank and are enclosed by thick gray lines.

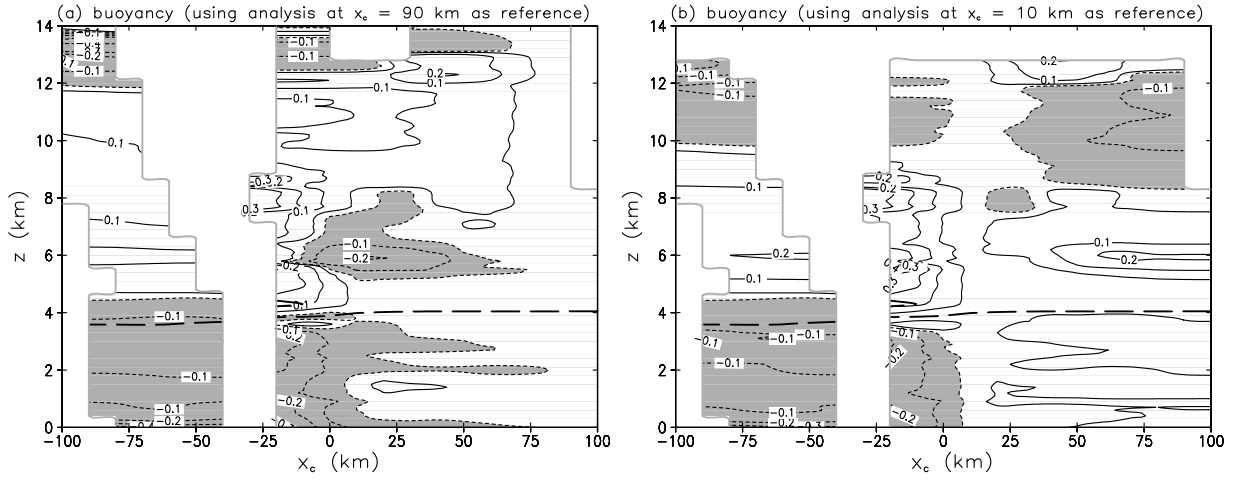


FIG. 19: The same as Fig. 18 except for buoyancy, where: (a) the profile at  $x_c = 90$  km is used as the reference and (b) the profile at  $x_c = 10$  km is used as the reference. Contour interval is  $0.1 \text{ m s}^{-2}$ , except intermediate contours of  $\pm 0.02 \text{ m s}^{-2}$  are included, and the zero contour is excluded. Values less than  $-0.02 \text{ m s}^{-2}$  are shaded. The melting level ( $T = 0^\circ \text{C}$ ) is shown as a thick dashed line.

Table 1: Rawinsonde launches on 15–16 May 2009 during VORTEX2. All rawinsondes were launched from a site 10 km south of Cherokee, Oklahoma, with the exception of S5 which was launched 6 km northwest of Cherokee, Oklahoma.

| ID | Time of<br>launch (UTC) | Time relative<br>to gust-front<br>passage (min) | Distance from<br>gust front at<br>launch (km) | Conditions<br>at launch                 |
|----|-------------------------|-------------------------------------------------|-----------------------------------------------|-----------------------------------------|
| S1 | 2138                    | +110                                            | +106                                          | Clear overhead                          |
| S2 | 2214                    | +74                                             | +71                                           | Edge of cirrus anvil overhead           |
| S3 | 2241                    | +47                                             | +45                                           | Thick cirrus overcast, edge of mammatus |
| S4 | 2307                    | +21                                             | +20                                           | Thick cirrus overcast, mammatus         |
| S5 | 2307                    | +4                                              | +4                                            | Overcast, shelf cloud to northwest      |
| S6 | 2339                    | −11                                             | −11                                           | Overcast, stratus, windy                |
| S7 | 0016                    | −48                                             | −46                                           | Overcast, stratus, light rain           |
| S8 | 0032                    | −64                                             | −61                                           | Overcast, stratus, light rain           |
| S9 | 0045                    | −77                                             | −74                                           | Overcast, stratus, light rain           |

Table 2: Observed values of cold pool depth ( $h$ , km) using VORTEX2 soundings S6–S8. The two rows list results using either S1 or S4 as the reference (i.e., environmental) sounding.

|                  | S6  | S7  | S8  |
|------------------|-----|-----|-----|
| S1 as reference: | 4.0 | 4.7 | 4.7 |
| S4 as reference: | 3.6 | 4.7 | 4.7 |

Table 3: Same as Table 2 but for cold pool intensity ( $C$ , m s<sup>-1</sup>).

|                  | S6   | S7   | S8   |
|------------------|------|------|------|
| S1 as reference: | 37.7 | 35.5 | 33.8 |
| S4 as reference: | 34.4 | 31.7 | 29.8 |