Deep Convection Initiation, Growth, and Environments in the Complex Terrain of Central Argentina during CACTI

ZHE FENG, ADAM VARBLE, JOSEPH HARDIN, JAMES MARQUIS, ALEXIS HUNZINGER, ZHIXIAO ZHANG, AND MANDANA THIEMAN

a Atmospheric Sciences and Global Change Division, Pacific Northwest National Laboratory, Richland, Washington
b Department of Atmospheric Sciences, University of Utah, Salt Lake City, Utah
c Science Systems and Applications, Inc., Hampton, Virginia
d NASA Langley Research Center, Hampton, Virginia

(Manuscript received 13 September 2021, in final form 15 December 2021)

ABSTRACT: This study characterizes the wide range of deep convective cloud life cycles and their relationships with ambient environments observed during the Cloud, Aerosol, and Complex Terrain Interactions (CACTI) field campaign near the Sierras de Córdoba (SDC) range in central Argentina. We develop a novel convective cell tracking database for the entire field campaign using C-band polarimetric radar observations. The cell tracking database includes timing, location, area, depth, merge/split information, microphysical properties, collocated satellite-retrieved cloud properties, and sounding-derived environmental conditions. Results show that the SDC exerts a strong control on convection initiation (CI) and growth. CI preferentially occurs east of the SDC ridge during the afternoon, and cells often undergo upscale growth through the evening as they travel eastward toward the plains. Larger and more intense cells tend to occur in more unstable and humid low-level environments, and surface-based cells are stronger than elevated cells. Midtropospheric relative humidity and vertical wind shear also jointly affect the size and depth of the cells. Rapid cell area growth rates exhibit dependence on both large environmental wind shear and low-level moisture. Evolution of convective cell macro- and microphysical properties are strongly influenced by convective available potential energy and low-level humidity, as well as the presence of other cells in their vicinity. This cell tracking database demonstrates a framework that ties measurements from various platforms centering around convective life cycles to facilitate process understanding of factors that control convective evolution.

SIGNIFICANCE STATEMENT: The purpose of this study is to develop a framework that ties coordinated radar, satellite, and radiosonde measurements around tracking convective storm life cycles to facilitate process understanding of atmospheric environments that control storm evolution. The processes coupling storm life cycles and local environments remain inadequately understood and are poorly represented in weather and climate models. Our results demonstrate the importance of atmospheric instability, low- and midtropospheric moisture, changes of wind with height, and interactions among nearby storms in affecting the formation and growth of convective storms. The storm database developed in this work enables future studies for comprehensive exploration of processes that lead to improved mechanistic understanding of storm evolution and their representations in models.

KEYWORDS: Convective clouds; Convective storms; Convective-scale processes; Storm environments; Cloud tracking/cloud motion winds; Radars/Radar observations

1. Introduction

Major mountain ranges of the world have significant influence on the initiation and organization of deep convection. Long-lived and intense mesoscale convective systems (MCSs) are primarily found downstream of mountain ranges within prominent baroclinic zones (Laing and Fritsch 1997, 2000). MCSs are responsible for producing the majority of the precipitation in these regions (Feng et al. 2021; Nesbitt et al. 2006), such as in the lee of the Rocky Mountains (Ashley et al. 2003; Feng et al. 2019; Fritsch et al. 1986; Haberlie and Ashley...
and the Andes (Durkee et al. 2009; Rasmussen et al. 2016; Romatschke and Houze 2013). Although the essential synoptic-scale ingredients for triggering deep convection and upscale growth to MCSs are largely known (Laing and Fritsch 2000; Rasmussen and Houze 2011; Song et al. 2019, 2021), our understanding of the processes coupling initiation and growth of convection with local environmental conditions remains inadequate, including internal cloud processes (e.g., evolution and structure of updrafts, precipitation, and cold pool formation) and local environmental interactions (e.g., entrainment–shear interactions with updrafts, cold pool interactions with the environment with dependence on depth and intensity). These processes are poorly represented in climate models, as manifested in the persistent warm and dry bias in the central United States (Klein et al. 2006; Lin et al. 2017; Morcrette et al. 2018) and in South America (Carril et al. 2012; Solman et al. 2013). Higher-resolution regional models improve the representation of convective processes by avoiding parameterization of deep convection, but still have persistent biases in convective updraft and downdraft intensities, structure and evolution of convective cells, and precipitation distributions (Fan et al. 2017; Hagos et al. 2014; Han et al. 2019; Varble et al. 2020, 2014a,b; Wang et al. 2020).

The relative importance of various factors that impact deep convection initiation (CI) and growth are difficult to examine in observations, partly because factors are correlated and interact across a variety of spatiotemporal scales. In addition, comprehensive observations of near-cloud ambient conditions leading up to CI and subsequent evolution of convection are very limited, owing to the difficulty to target the precise location and timing of CI in a highly heterogeneous mesoscale environment. Several past field campaigns have made use of terrain features that focus convective activity to better understand cumulus and deep convection life cycles, including CuPIDO (Damiani et al. 2008), COPS (Wulfmeyer et al. 2008), and CISP (Browning et al. 2007). Despite these efforts, sample sizes under a variety of environmental conditions remain limited, hindering a robust understanding of CI and growth processes and verification of these processes in models. Some of the world’s most intense deep convection is observed in central Argentina (Zipser et al. 2006), where large MCSs dominate the total and extreme rainfall (Rasmussen and Houze 2011; Rasmussen et al. 2016). Located downstream of the Andes, central Argentina produces a high frequency of orographically generated cumulus clouds over the Sierras de Córdoba (SDC) range that rises 2000 m above the surrounding plains. These clouds often grow to initiate deep convection and organize into MCSs (Mulholland et al. 2018; Rasmussen and Houze 2011). Convective cloud evolution is also influenced by synoptic systems in the region including (i) the northwestern Argentinean low (Seluchi et al. 2003), (ii) northward-moving cold fronts from the Patagonia region (Seluchi et al. 2006), and (iii) the South American low-level jet (Montini et al. 2019), which transports warm and moist air from the Amazon into the region (Borque et al. 2010; Salio et al. 2007). Interactions between such synoptic features and the SDC induce upslope flow and low-level ascent (Marquis et al. 2021; Singh et al. 2022), facilitating frequent CI and often rapid upscale growth between the spring and fall seasons (Cancelada et al. 2020; Mulholland et al. 2018). Associated terrain-induced mesoscale variations in convective environmental ingredients affect the precise locations of CI, subsequent intensification of deep convection, and rapid upscale growth to MCSs. For example, Marquis et al. (2021) and Singh et al. (2022) illustrate that CI is favored where SDC-induced flow reduces CIN and mitigates entrainment-driven dilution of clouds by locally increasing low-level moisture. Trapp et al. (2020) hypothesize that a rapid intensification to severe thunderstorms may occur within pockets of large wind shear in the wake of the SDC. Mulholland et al. (2019, 2020) further demonstrate the role of specific SDC terrain structure in simulations in altering vertical wind shear, CAPE, CI timing, and convective morphological evolution via upslope flow, mountain waves, and blockage of cold pools. However, these links between the behavior of deep convection and local environmental factors are based on case studies and climatologies using reanalyses, leaving open the question of their applicability across a wide spectrum of convective conditions.

Many previous studies that examine the environments associated with deep convection in South America emphasize large MCSs or extreme convective events, such as those containing very deep or wide convective cores (Anabor et al. 2008; Rasmussen and Houze 2011, 2016; Rasmussen et al. 2016; Salio et al. 2007) or supercells (Borque et al. 2020; Mulholland et al. 2019, 2018; Trapp et al. 2020). However, many deep convective clouds are much less extreme, but collectively represent large radiative and hydrological impacts on regional and global climate. A broader understanding of the entire range of environmental controls on the deep convective life cycle is needed, particularly using a comprehensive set of coordinated in situ and remote sensing measurements that is consistent across a large sample of cases, to isolate and quantify specific processes that result in improved parameterizations and predictions by weather and climate models.

To improve understanding of two-way interactions between convective clouds and their environments, the Cloud, Aerosol, and Complex Terrain Interactions (CACTI) field campaign was conducted to observe orographic clouds occurring in a wide range of environmental conditions over the SDC between October 2018 and April 2019 (Varble et al. 2021). This campaign overlapped with the Remote Sensing of Electrification, Lightning, and Mesoscale/Microscale Processes with Adaptive Ground Observations (RELMAPAGO) field campaign (Nesbitt et al. 2021). A key objective of CACTI and RELMAPAGO was to better understand processes that promote or suppress CI, subsequent upscale growth of convection, and mesoscale organization. An Atmospheric Radiation Measurement (ARM) Mobile Facility (AMF; Mather and Voyles 2013) and the C-band Scanning ARM Precipitation Radar (C-SAPR2) were deployed at the same site during the 6.5-month campaign period. Matched with high-resolution geostationary satellite observations, the campaign successfully measured a multitude of CI
and convective growth events along with their near-cloud atmospheric conditions (Varble et al. 2021).

The goal of this study is to characterize the range of deep convective clouds observed by the C-SAPR2 radar and use these data to examine the relationships between local environmental conditions, CI, and subsequent growth. To achieve this goal, we develop a novel convective cell tracking database (Feng et al. 2022) that utilizes measurements of mean convective cell motion between consecutive routine radar three-dimensional volume scans. Radiosonde-derived environmental conditions and geostationary-satellite-retrieved cloud properties are matched to the resulting cell tracking database, allowing statistical examination of the relationships between environments and deep convection evolution over a large sample of cells.

This paper is organized as follows. Section 2 describes the observation datasets used. Section 3 introduces the methodology for the identification and tracking of convective cells, as well as matching with satellite and sounding measurements. Section 4 presents the spatiotemporal characteristics of the convective cells observed during the campaign. Section 5 discusses relationships between local environmental conditions and convective cell evolution, relates them to hypotheses posed in the recent literature regarding the environmental governance of deep convection in the region, and poses an additional hypothesis regarding interactions between convective cells during CI and subsequent growth. Finally, section 6 summarizes results and conclusions.

2. Datasets

a. C-SAPR2 radar dataset and processing

During the CACTI field campaign, the C-SAPR2 performed a regular 15-min repeating scan sequence that started with a 15-tilt plan position indicator (PPI) volume scan between 0.5° and 33° elevations (Hardin et al. 2018), followed by two sets of six hemispheric (horizon-to-horizon) range–height indicator (HSRHI) scans (Hardin 2018) every 30° in azimuth (see Fig. S1 in the online supplement). The radar operated around the clock in this mode from mid-October 2018 until the end of February 2019. Occasional mechanical issues resulted in downtime from late December to January 20 and parts of February. After 2 March 2019, PPI scans became no longer possible given mechanical problems, and thus this period is not included in this study.

The C-SAPR2 data were calibrated using a combination of on-site engineering measurements and cross-instrument comparisons (Hardin et al. 2020a). The Taranis radar processing package (Hardin et al. 2020b) was used to process the data. First it removes clutter, second trip echoes, and other nonmeteorological returns. Then the package calculates specific differential phase ($K_{DP}$), which is used for radar reflectivity ($Z$) and differential radar reflectivity ($Z_{DR}$) attenuation correction based on a hybrid self-consistency–linear programming algorithm. Last, the package retrieves quantitative precipitation estimates (QPE) and rainwater contents following Brigni and Chandrasekar (2001), raindrop mass-weighted diameters ($D_m$) following Matrosov et al. (2005), and hydrometeor identifications using the CSU RadarTools package (Lang et al. 2019). The Taranis-processed data are then converted to a Cartesian grid from their native radial coordinate using PyART (Helmus and Collis 2016). The Cartesian grid spacing is 500 m in the $x$, $y$, and $z$ directions, covering a region with 220-km diameter and depth up to 20 km above the surface (~1141 m MSL).

b. AMF sounding dataset

Radiosondes were launched at 1200 (0900 LT), 1500, 1800, 2100, and 0000 (2100 LT) UTC at the AMF site on most days when deep convection was forecasted. On other days, radiosondes were launched at 0000, 1200, 1600, and 2000 UTC. Slightly more than 50% of days had five launches per day during C-SAPR2 operations. This launch strategy focused on the daytime environmental evolution east of the SDC ridge that was most relevant to orographic clouds. This sounding dataset was synthesized into the ARM Interpolated Sounding (INTERPSONDE) value-added product (Fairless and Giangrande 2018), which linearly interpolates atmospheric state variables in time at each height level to 1-min resolution and constrains the profiles between radiosonde launch times with AMF microwave-radiometer-retrieved column integrated water vapor. In this study, we utilize INTERPSONDE data at times matched to the 15-min C-SAPR2 PPI volume frequency to characterize the near-cloud environment.

We evaluate convective available potential energy (CAPE), convective inhibition (CIN), level of neutral buoyancy (LNB), initial height of a lifted parcel starting level (LPL), and level of free convection (LFC), each for the most unstable (MU) parcel in the lowest 4 km of the atmosphere (i.e., the parcel with the largest CAPE), following Nelson et al. (2021). Other environmental metrics evaluated include water vapor mixing ratio and relative humidity (RH) at 850, 700, and 500 hPa, and 0–3-, 0–6-, and 0–9-km bulk vertical wind shear.

c. GOES-16 cloud property retrievals

Cloud property retrievals from the Geostationary Operational Environmental Satellite 16 (GOES-16) were produced using the NASA Langley Satellite Cloud Observations and Radiative Property retrieval System (SatCORPS) algorithm (Minnis et al. 2021; Trepte et al. 2019; Yost et al. 2021). This dataset (Smith and Thieman 2019) provides various cloud properties at 2-km and 15-min spatiotemporal resolution over an approximately 1450 km × 1350 km region centered at the AMF site. Retrievals include cloud-top temperature (CTT), height, pressure, cloud phase (liquid or ice), and column integrated quantities such as cloud optical depth, and liquid/ice water path. For this study, we focus on CTT associated with convection observed within the C-SAPR2 coverage (110-km radius from the AMF site). We parallax correct the GOES-16 retrievals to match them (in space and time) to the C-SAPR2 gridded data using the nearest-neighbor method from xESMF (Zhuang 2020).
3. Methodology

3a. Convective cell identification

Previous studies often use the constant-altitude plan position indicator (CAPPI) at a low altitude (e.g., 2 km) above the surface to identify convection from ground-based scanning radar observations (e.g., Hagos et al. 2014; Powell et al. 2016; Varble et al. 2014a). To mitigate significant terrain blockage west of the radar site, we instead use composite reflectivity, defined as the maximum radar reflectivity in a column, and start the column 500 m above the terrain elevation to be certain that no ground clutter contaminates signals. The composite reflectivity also increases the probability of detecting cells in their earliest stages of formation because ice and radar-detectable hydrometeors aloft precede the presence of lower-level precipitation.

We use a modified horizontal radar reflectivity texture algorithm based on Steiner et al. (1995) to identify convective cells. This algorithm utilizes the horizontal “peakedness” of radar reflectivity (i.e., the difference between a grid point reflectivity and its surrounding background reflectivity) produced by convective echoes. Since the goal of this study is to characterize the initiation and growth of individual convective cells rather than to identify large regions constituted by numerous cells, we made the following changes to the Steiner algorithm. First, we set all points that are less than 0 dBZ to 0 dBZ. Then a convective grid point is identified as one that satisfies either of these conditions:

1) composite reflectivity \( Z > 60 \text{ dBZ} \), or
2) difference between composite \( Z \) and background composite \( Z_{bkg} \) (defined as the mean \( Z \) within an 11-km radius centered at the grid point in the radar domain) is

\[
\Delta Z = \begin{cases} 
10 \cos \left( \frac{\pi Z_{bkg}}{2 \times 60} \right), & Z_{bkg} \geq 0 \text{ dBZ} \\
0, & Z_{bkg} > 60 \text{ dBZ} 
\end{cases}
\]

A convective core is identified as horizontally contiguous convective grid points if they constitute an area of at least 4 km\(^2\) (16 grid points). After convective cores are identified, they are expanded outward into surrounding grids using a \( Z_{bkg} \)-dependent radius step function (see Fig. 6 in Steiner et al. 1995) to define convective cells that consist of core area and noncore area. The radius step function is modified from Steiner et al. (1995) to start from 1 km at \( Z_{bkg} = 25 \text{ dBZ} \) with a 0.5-km radius increment per 5-dB \( Z_{bkg} \) increase out to 5 km at \( Z_{bkg} = 60 \text{ dBZ} \). Each convective cell is then assigned a unique cell number. Last, to facilitate easier cell tracking, the area of each unique convective cell (cell masks) is incrementally expanded outward within a 5-km radius with 1-km-radius steps, starting from the largest cell in a scene without merging of cells. This expansion creates larger footprints for the identified convective cells to improve cell tracking accuracy (section 3c) and is performed at each scan time.

Our definition of convective cores is designed to better separate individual convective cells within “convective complexes” than the Steiner et al. (1995) algorithm, which focused more on classifying convective and stratiform precipitation regions. We increased the \( Z \) and peakedness thresholds for convective cores to further avoid misidentification of convection in stratiform radar brightband in our composite reflectivity metric. Furthermore, including echo-free locations in the calculation of \( Z_{bkg} \) increases identification of relatively weak and isolated convective cells. Finally, our 500-m gridded radar dataset is substantially higher resolution than the 2-km grid in Steiner et al. (1995), which required threshold adjustments in the algorithm such as the \( Z_{bkg} \)-dependent radius step function.

Figure 1 illustrates the convective cell identification results in relatively weak, moderate, and intense convection. Our modified algorithm successfully identifies individual weak convective cells and moderate intensity cells within larger convective regions and complexes. Further, it identifies cells located in strong stratiform rain brightband signals without identifying the brightband (e.g., ~40-dBZ brightband areas between 60- and 110-km radius south and southeast of the radar at 0500 UTC 27 November 2018, Figs. 1c,d). HSRHI scans illustrate the vertical structure of convective cells of various depths that correspond to the convective cells identified in the PPI scans.

3b. Convective feature advection estimates

It is challenging to accurately track small and/or fast-moving convective cells occurring in strong background winds between 15-min radar volume intervals. To assist cell tracking, we developed a methodology that includes an estimate of the mean advection velocity of convective features, which is illustrated with an example in Fig. 2. This approach is similar to previous studies of Lagrangian cell tracking that use storm movement from previous times to “predict” the locations of convective cells in subsequent times (Dixon and Wiener 1993; Heikenfeld et al. 2019; Moseley et al. 2013).

First, significant radar echoes are identified using composite reflectivity > 10 dBZ. Second, we perform a two-dimensional cross correlation between composite reflectivity fields within significant echoes from two adjacent 15-min scans. Offsets in the \( x \) and \( y \) directions for every 15-min scan are determined. Each convective cell masks between times and improves the accuracy of tracking individual cells, which will be discussed in more detail next.

In practice, we calculated a time series of domainwide advection distance in both the \( x \) and \( y \) directions for every pair of consecutive 15-min radar scans. For radar scans lacking sufficient convective echoes (significant echo occupying < 10 grid points), the advection estimate is set to 0. After calculating the advection time series, a 9-point (2-h) median filter is applied to remove occasional bad advection estimates.
The resulting advection distance time series is then supplied to the cell tracking procedure.

c. Convective cell tracking

To track convective cells, we modify the Flexible Object Tracker (FLEXTRKR; Feng et al. 2018; Wang et al. 2020) algorithm to include our cell advection velocity estimates. Expanded convective cell area masks (described in section 3a) are shifted between consecutive 15-min scans based on the advection distance estimated in section 3b to improve tracking via time overlap of convective objects. Cells whose area masks spatially overlap > 30% between consecutive radar scans are tagged as the same cell and are further tracked between additional consecutive radar volumes.

If multiple cells from a previous scan overlap with one cell at the next time step, the largest one from the previous time step is continued to the next time step, and the smaller cell track is cataloged as terminating in a cell merger. Cell “splitting” is handled similarly, though in a reverse direction—if two cells overlap with one from a previous time, the larger of the pair is matched with the previous single cell and the smaller of the pair begins a new track that results from a cell split. Statistics of cell characteristics during their full tracks are saved, including centroid location, identification number,
lifetime, area, merger and splitting times (if applicable), maximum radar echo top heights, minimum GOES-16-retrieved CTT, and vertical profiles of maximum reflectivity, \( Z_{DR} \), rainwater content, rain rate, and \( D_m \). Cell area and track number masks are written back to the native 500-m-resolution Cartesian-gridded radar data to facilitate matching with other geolocated datasets.

An example of a tracked CI and upscale growth event is shown in Fig. 3. Convective cells were first detected along the SDC ridgeline just west of the C-SAPR2 radar between 1700 and 1800 UTC 25 January 2019. Several cells quickly intensified and were nearly stationary during their first 1–2 h (Figs. 3a–c). After 1900 UTC, a group of intense cells formed to the southeast of the radar. These cells merged into a large convective complex (Figs. 3d,e), which traveled northeastward between 2000 and 2100 UTC. The complex developed into a convective line that merged with the original intense long-lived cells at \( \sim 2115 \) UTC before moving out of range of the radar after 2200 UTC.

Cell track center locations correspond well with radar reflectivity peaks consistent with individual storms (Fig. 3). Occasional large “jumps” in the cell center location are typically associated with cell merger or splitting events (e.g., the cell track annotated in Figs. 3d,e,h,j). A different event containing many faster-moving cells that both initiated within the radar domain and entered it from the outside is shown in supplemental animation S2. Most of the cells are tracked quite well in that example, suggesting that our algorithm works well for both slower-moving and faster-moving cells.

We apply this FLEXTRKR convective cell tracking algorithm continuously for the entire field campaign period when the C-SAPR2 radar collected 15-min PPI volume data. Tracking is terminated and restarted when encountering radar data gaps exceeding one missing PPI data volume. A total number of 6895 individual cells were tracked during the campaign period.

An example of paired environmental and cell metrics for the 25 January 2019 event is shown on the inset of Fig. 3k. The rapid intensification of this cell occurred in an environment with very high MUCAPE (7169 J kg\(^{-1}\)) and 23 m s\(^{-1}\) bulk 0–6-km vertical wind shear. Schumacher et al. (2021) show that this event had the highest CAPE values observed by radiosondes during the 2018/19 warm season, while the shear was common. Studies using reanalysis derived CAPE values show that 7000 J kg\(^{-1}\) CAPE values may be limited to occurring once per year or less in the SDC region (Bruick et al. 2019; Mulholland et al. 2018), but a lack of long-term radiosonde data during afternoon hours limits quantifiable climatological context. The cell achieved a minimum CTT of \(-93^\circ\)C, a maximum radar reflectivity of 68 dBZ, and lasted 3.75 h before merging into a convective line to form an intense MCS. Because each tracked cell contains environmental metrics and storm properties observed by radar and satellite, our database enables case studies of individual events (e.g., Fig. 3) as well as broader statistical analyses that contextualize one another and lead to better understand of processes that control the life cycle of deep convection.

4. Spatiotemporal characteristics of convective cells

a. Terrain impact on convective cell evolution

Widely variable MUCAPE, 850-hPa water vapor mixing ratio, and most unstable LNB (MULNB) were observed during the campaign (Fig. 4a); 850-hPa water vapor mixing ratio and MULNB often covary, and relatively large values of each correspond to large values of parcel buoyancy (e.g., MUCAPE > 1000 J kg\(^{-1}\)). These high MUCAPE and low-level moisture conditions are often associated with north to northeasterly low-level flow, sometimes in the form of a low-level jet that transports heat and moisture from the Amazon (Varble et al. 2021).
FIG. 3. Example cell tracking for a convective upscale growth event on 25 Jan 2019. (b)–(i) Each panel represents a 30-min interval. Individual cell tracks are shown by symbols connected with black lines. Symbols denote tracked cells and are color coded by their lifetime (color bar at upper left of each panel). CI locations are indicated by a larger symbol for each track. (bottom) Time evolution of (j) tracked cell area and (k) minimum cloud-top temperature for cells lasting longer than 1 h are shown. Symbols in (j) represent cell merge/split status recorded and selected environmental parameters and statistics for cell track 6315 initiated at 1700 UTC northwest of the radar is shown in the annotation in (k). Two long-lived cell tracks 6315 and 6321 are marked in (a)–(i) and their temporal evolution are shown in (j) and (k). An animation of this event is provided in the supplemental animation S1.
There are multiple periods with continuous or near-continuous convective cell activity (Fig. 4b) that correlate with large MUCAPE and MULNB, and ample low-level moisture (850-hPa water vapor mixing ratio \(\geq 8\) g kg\(^{-1}\)), which occur more frequently under northerly and northeasterly low-level winds (not shown). Some events result in large and deep convective cells, as indicated by the maximum cell diameter larger than 15 km ([\(\sim 177\) km\(^2\)], and 20 dBZ echo-top heights (ETHs) reaching 12–15 km AGL (Figs. 4c,d). There are a few events with extremely deep cells (e.g., 20-dBZ ETH reaching 18–20 km AGL on 11–12 November 2018, 14 December 2018, and 25 January 2019). These events occur in very unstable conditions where peak MUCAPE exceeds 3000 J kg\(^{-1}\) and peak MULNB exceeds 14 km AGL.

In general, AMF sounding observations indicate that low-level moisture, MUCAPE and MULNB are higher in the summer (January–February 2019) than spring (October 2018), with lower LPL, weaker 0–6-km wind shear and slightly higher midtropospheric RH (Fig. S2). Consistent with these local environmental differences, convective cells in the summer are wider, deeper, and have higher maximum reflectivity (Fig. S3) than those in spring. During November–December 2018, the

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**Fig. 4.** Time series of the cell tracking database during the entire CACTI field campaign period when C-SAPR2 performs PPI scans. (a) Select sounding parameters, (b) number of cells initiated in the entire domain (black), east of the primary SDC ridgeline (orange), and west of the SDC ridgeline (blue), (c) distributions of lifetime maximum cell equivalent diameter \(D = \sqrt{\text{Area}/\pi}\), (d) distribution of lifetime maximum 20-dBZ echo-top height, and (e) domain-mean total rain rate (solid lines) east of the SDC ridgeline (black) with convective (red) and stratiform (blue) components and their associated rainfall accumulation amount (dash lines). Data are plotted in 6-hourly intervals. Gray color fill in (b)–(e) indicates periods when C-SAPR2 did not collect data. Vertical black dash lines denote the first day of each month.
near-storm environments and convective cell characteristics contain generally intermediate values between October and January–February. Varying synoptic patterns from spring to summer likely contribute to these local environments and convective cell evolution. Disentangling the relative importance of environmental factors at different scales to CI and upscale growth is beyond the scope of this work, but the wide variety of environmental conditions and convective cell characteristics captured in the cell tracking database enables such studies.

CI most frequently occurs just east of the SDC ridgeline (∼64.9°W), with a secondarily enhanced CI frequency just west of the ridge (Fig. 5a), illustrating that the SDC range exerts a direct control on CI. CI events occurring east of the primary SDC ridgeline account for 2.5 times more tracked cells (4918 cells) than those to the west of it (1969 cells) (Fig. 4b). Primary cell tracks end slightly east of the primary initiation region, indicating many cells do not travel far before dissipating (Fig. 5b). However, a secondary peak in cell track ending location occurs at the far eastern edge of the radar coverage domain, suggesting a portion of the cells survive longer and travel a much greater distance out onto the plains. The cell lifetime distribution shows that most of the tracked cells last 15–30 min, with an exponential decrease in the number of longer-lived cells (Fig. S4). However, about 50% of cells last beyond the lifetime of a typical single cell (Markowski and Richardson 2010; Wilhelmson and Chen 1982), and ∼30% (1562 cells) last an hour or more, suggesting they are more organized, and some could develop upscale to mesoscale complexes.

A clear contrast of cell area and depth is seen between cells occurring west and east of the SDC (Figs. 5c,d). Cell area and depth both are largest east of the SDC. The deepest cells tend to develop between 20 and 60 km east of the SDC ridgeline, while the widest cells occur slightly farther east of this location. With typical eastward cell movement and initiation locations, these results indicate rapid deepening and intensification on the eastern slopes of the SDC, followed by widening in area as cells grow rapidly upscale en route to the eastern plains. The prevalence of deep and/or wide convective cells near and east of the SDC is consistent with satellite observations from the Tropical Rainfall Measuring Mission (Rasmussen and Houze 2011). Further, increased cell area into the afternoon and evening is consistent with upscale growth of cells into MCSs, which is common in the evening hours (Mulholland et al. 2018; Zhang et al. 2021).
b. Diurnal cycle of convective cells

A trimodal peak in the diurnal cycle of total cell initiation number is observed, including an early afternoon peak followed by a stronger nocturnal peak, and a weak early morning peak (Fig. 6a), consistent with Mulholland et al. (2018). For cells that initiated east of the SDC ridgeline, a pronounced afternoon peak occurs at 1500–1700 LT (Fig. 6b) that contributes to ~75%–80% of the afternoon total cell initiation count (Fig. 6c). Substantial increases in cell maximum area, ETH, and maximum reflectivity coincide with this peak in afternoon convective activity. After ~1700 LT, the number of CI events decreases to a relative minimum at ~1900 LT; however, cell maximum area, ETHs, and maximum reflectivity continue to increase into the evening hours, with 75th–90th-percentile values peaking in the evening and only slightly diminishing until 0300 LT. The number of CI events increases again starting at ~2000 LT. This secondary peak of CI events occurs at around midnight before precipitously dropping at around 0300 LT when area, ETH, and maximum reflectivity also drop. The diurnal cycle of thermodynamic environmental INTERPSONDE measurements suggests that MUCIN erodes and typically reaches a minimum at around the afternoon peak in cell initiation (Varble et al. 2021). MUCAPE peaks a bit later, coinciding with increased cell area, depth, and intensity in the evening. Evening convection typically becomes more based in elevated CAPE (Varble et al. 2021), potentially sustained by low-level jet inflow that delays the decay of tracked cell number, depth, area, and intensity. The cause for the third peak in CI at ~0800 LT east of the SDC ridgeline is currently unclear.

Preferential CI location shifts with the diurnal cycle (Fig. 7). During the day, particularly in the afternoon hours, CI is most frequently concentrated within 25 km east of the SDC ridgeline (Figs. 7b,c). The afternoon peak in CI is likely associated with the orographic upslope flow east of the SDC (Marquis et al. 2021; Nelson et al. 2021). During days with strong diurnal heating, solar insolation often allows mountains to act as sources of elevated heat and instability where convective clouds and storms first occur (Kirshbaum 2011, 2013). Preferential locations also shift from over the northern SDC in early afternoon to the central SDC later in the afternoon. This may be related to the elevated plateau-like topography of the northern SDC as compared to the sharper ridgeline of the central and southern SDC creating differences in the daytime growth of the boundary layer and orographic flows. After sunset, the most prominent CI location transitions to the steep western slopes of the SDC, which remains a favored CI location through the night (Figs. 7e–h), consistent with Cancelada et al. (2020). CI becomes slightly more widespread after midnight, including areas east of the ridgeline (e.g., 0300–0600 LT; Fig. 7f). The nocturnal CI peak may be
partially related to complex interactions between the SDC and the South American low-level jet, which increases in strength at night as it decouples from the boundary layer (Zhang et al. 2021). Besides diurnal changes in low-level jet–terrain interactions, the cessation of thermal upslope flow may alter preferential regions of convergence and lift that support CI. In addition, lift provided by cold pools associated with nocturnal MCSs may also promote CI. These processes require further research beyond the scope of this study.

5. Environmental controls on convective cells

Our large population of 6895 tracked cells offers a unique opportunity to statistically evaluate the impact of various environmental factors on CI and subsequent growth.

a. Environmental impacts on maximum cell area, intensity, and depth

Several interesting relationships among cell characteristics are highlighted with joint probability density functions (PDFs; Fig. 8). In general, the deepest cells tend to have the highest intensity in terms of maximum radar reflectivity (Fig. 8b). Further, the widest cells tend to be the most intense (Fig. 8a) and deepest (Figs. 8c,d). There is substantial variability in cell intensity and depth for a given cell area, particularly for smaller cells (Figs. 8a,c,d); however, the range of cell intensities varies even more so per cell depth (Fig. 8b). There is a bimodal distribution of minimum CTT and maximum radar reflectivity (Fig. 8b). A relatively shallow cell mode consists of cloud tops around the freezing level (0° to −10°C) with modest peak reflectivities of 25–40 dBZ, while a more prominent deep cell mode has cloud tops most commonly between −45° and −60°C with peak reflectivities between 35 and 55 dBZ. A nonnegligible number of deep cells have minimum CTTs lower than −75°C or peak reflectivity exceeding 60 dBZ.

Relationships between environmental parameters and cell characteristics are evaluated using a similar joint PDF framework (Fig. 9). We focus on the 4918 cells that initiate east of the SDC ridgeline because they account for over 70% of the total cell tracks and have environments better characterized by the 3-hourly radiosonde data launched at the AMF site (Fig. 6). We note that environments are likely less accurately depicted by the INTERPSONDE data for cells that initiate relatively far from the AMF site, although most cells initiating east of the SDC ridgeline are located within ~70 km of the AMF site along the eastern SDC slopes.

Three of the examined environmental thermodynamic parameters show consistent relationships with cell area and intensity. Higher MUCAPE (Fig. 9a), MULNB (Fig. 9b), and 850-hPa water vapor mixing ratio (Fig. 9c) correspond to wider and stronger cells. These dependencies are largely related to wider cells being more intense. For a given maximum reflectivity, changes in cell area do not correspond to large changes in MUCAPE, MULNB, or 850-hPa water vapor, but for a given maximum cell area, increasing maximum reflectivity corresponds to significant increases in those environmental parameters. A higher MULNB allows cells to achieve a greater depth and potentially favors more intense convection. Varble et al. (2021) noted that MUCAPE and MULNB most often peaked in the early evening (1700–1800 LT), while MUCIN usually reached a minimum earlier in the afternoon (~1500 LT). MUCIN for our tracked cells is consistently low (typically <20 J kg⁻¹) regardless of cell characteristics (Fig. S5), suggesting MUCIN has been eroded leading up CI. A similar relationship between 850-hPa water vapor mixing ratio and MUCAPE/MULNB with cell area/intensity.
(Fig. 4a) suggests that convective potential and intensity are regulated to a first order by changes in low-level moisture. The most unstable lifted parcel starting level (MULPL) primarily varies with cell maximum intensity rather than cell area (Fig. 9d). As lifted parcels become increasingly more surface based (lower MULPL), convective cells become more intense. Elevated cells that more often initiate in overnight and morning hours (Schumacher et al. 2021) tend to be weaker than cells that initiate during the afternoon and early evening, consistent with the results shown in Fig. 6c and higher MUCAPE, MULNB, and 850-hPa water vapor in Figs. 9a–c for lower MULPL (surface and boundary layer based) cells.

Midlevel RH and deep-layer vertical wind shear also affect the size and depth of convective cells. Though higher 500-hPa RH generally favors larger and deeper convective cells, more nuanced relationships also exist between them (Fig. 9e). For a given cell area (x axis), higher RH favors deeper cells (y axis). This suggests that a drier midtroposphere may enhance entrainment-driven dilution of updraft buoyancy and limit the vertical growth of cells, consistent with recent theoretical and idealized modeling studies (Morrison et al. 2020; Peters et al. 2020). On the other hand, for cells that reach a given depth, smaller cells occur in higher-RH environments, suggesting narrow cells may only reach a certain depth if RH is high enough to limit buoyancy dilution; whereas relatively wide cells may be comparatively less prone to entrainment-driven buoyancy dilution (Morrison 2017). Last, larger cells generally occur in stronger 0–6- km bulk vertical wind shear environments, except for very deep cells with CTTs below −60°C (Fig. 9f). However, sample sizes for these very cold cloud tops are limited such that relationships with environmental parameters may be strongly influenced by individual events (∼20–40 per bin, Fig. 8d). Interestingly, for cells with minimum CTT warmer than −60°C, stronger shear is associated with wider, but shallower cells and weaker shear with narrower, but deeper cells. This suggests weaker shear allows narrow updrafts to develop taller while stronger shear results in wider updrafts. Similar relationships with 0–3- and 0–9-km bulk vertical wind shear are found (not shown). The limiting of cell depth in environments with stronger wind shear may be consistent with suppression of updrafts by shear-induced
downward-oriented pressure gradient accelerations, as demonstrated by Peters et al. (2019). However, that study focuses on isolated cloudy updrafts, whereas our dataset also incorporates cells within mesoscale organized convective systems that complicate interpretations of causes for relationships. We note that the effect of 500-hPa RH (Fig. 9e) is superimposed on vertical shear (e.g., higher RH correlates with lower shear); therefore, these factors may be controlling each other’s cell area and depth relationships. In addition, MUCAPE, MULNB, and 850-hPa water vapor slightly increase as maximum cell area increases for cells reaching a given minimum CTT (Fig. S6), which indicates they may also contribute to RH and shear correlations with cell width and depth.

The impacts of MUCAPE, 850-hPa water vapor mixing ratio, and 0–6-km bulk vertical wind shear at the time of CI on cell area growth rate are shown in Fig. 10. Results show that these environmental metrics only loosely correspond to

![Fig. 9](image1.png)

**Fig. 9.** Relationships between mean sounding-computed environmental properties at cell initial time and cell characteristics. (a) Mean MUCAPE, (b) mean MULNB, (c) mean 850-hPa water vapor mixing ratio, and (d) mean MULPL as functions of maximum cell area and reflectivity, and (e) 500-hPa relative humidity and (f) 0–6-km bulk vertical wind shear as functions of maximum cell area and minimum cloud-top temperature. Only cells that initiate east of SDC ridge line (64.9°W) are included.

![Fig. 10](image2.png)

**Fig. 10.** Violin plots of cell area growth rates as functions of (a) MUCAPE, (b) 850-hPa water vapor mixing ratio, and (c) 0–6-km bulk wind shear. Mean growth rates in low, medium, and high groups are plotted with magenta horizontal lines and their values are shown in the legend. Distributions that are statistically significant (Kolmogorov–Smirnov test p value < 0.05) vs other groups are marked with a check mark; otherwise, they are marked with a cross. Sample size in each group is shown above the violin plots. Growth rate is calculated by dividing the cell area difference (between cell maximum area and area at CI) by the time it takes to reach maximum area. Only cells initiated east of the SDC ridgeline with lifetimes of at least 1.75 h are included.
cell area growth rates. The most obvious distinctions lie at opposing ends of the environment distribution; cells occurring in the greatest MUCAPE (Fig. 10a), 850-hPa humidity (Fig. 10b), and 0–6-km shear (Fig. 10c) exhibit clearly faster growth compared to the environments with the lowest values. However, growth rates in moderate environments are not always significantly different from lesser or greater environments depending on the metric. For example, growth rate differences between moderate (400–1000 J kg$^{-1}$) and high (>1000 J kg$^{-1}$) MUCAPE are not statistically significant (Kolmogorov–Smirnov test $P$ value > 0.05). Similarly, low and moderate values of wind shear correspond to statistically similar cell growth rates. These findings suggest that additional factors affect cell growth rates, potentially including synoptic and mesoscale circulations, low-level moisture flux, cold pool–wind shear interactions, and gravity waves. Furthermore, interactions among neighboring cells (e.g., merging and splitting) may also play an important role on growth rate. The effects of cell interactions on their characteristics are further examined in section 5c.

b. Moisture and CAPE impacts on cell vertical structure evolution

Because the lifetime maximum cell area and intensity exhibit close relationships with 850-hPa water vapor mixing ratio and MUCAPE (Fig. 9), we further examine the effects of these two environmental factors on the evolution of vertical cell structure.

Figure 11 illustrates significant changes in composite vertical profiles of six important cell macro- and microphysical characteristics at CI time as a function of the 850-hPa water vapor mixing ratio. Cells that initiate in environments containing greater low-level moisture produce clearly greater radar reflectivity throughout the troposphere with cell horizontal area (the 20-dBZ echo area) that is significantly wider above the boundary layer (Figs. 11a,b). Though this correlation between environmental moisture and cell area occurs primarily at $z > 3$ km MSL. Cells at CI are deeper (Fig. 11a) as low-level moisture increases, and initial hydrometeors forming higher aloft will take longer to reach low levels, which may cause differences in maximum reflectivity to disappear at low levels at this initial cell time. The magnitude of this effect may also be impacted by the 15-min radar scan cycle.

In moderately to significantly moist environments, the cell maximum $K_{DP}$ profiles at CI time tend to be slightly greater above 6 km, and the tail values of $K_{DP}$ are also larger below 4 km (Fig. 11c), although the overall $K_{DP}$ magnitudes are
quite small as expected at CI. Similarly, larger Z_{DR} profiles throughout the column are observed in more humid environments (Fig. 11d). Rain rate profiles are similar among the three groups, although the tail values are still largest for cells initiated in moist conditions (Fig. 11e). Clear contrast is seen in raindrop diameter (D_{m}) profiles, consistent with Z_{DR} profiles where the D_{m} retrievals are based on power-law estimators derived from disdrometer measurements following the method from Bringi et al. (2003). Together, these microphysical signatures suggest that cells initiating in more humid low-level environments become initially deeper with larger raindrops at low levels, but rain rates are comparable. Larger raindrops may be caused by greater collision–coalescence due to greater moisture supporting faster precipitation formation with all else equal.

To further examine the impact of CAPE on the statistics of cell evolution, we show the time evolution of cell composite vertical profiles (median values) as a function of MUCAPE in Fig. 12. In this analysis, we eliminate the numerous short-lived cells (Fig. S2) and only consider those with a lifetime between 1.75 and 2 h that do not begin as splits or end as mergers. We further decompose this sample into three groups, with low (50–400 J kg^{-1}), moderate (400–1000 J kg^{-1}), and high (1000–7000 J kg^{-1}) MUCAPE with each group having between 26 and 39 cells.

Like the cell profiles in Fig. 11, clear differences are seen for both cell macro- and microphysical characteristics. Cells developing in increasingly unstable environments grow deeper, produce more intense reflectivity throughout the troposphere, and have larger areas (using area of the 30 dBZ reflectivity contour) in the mid- to upper troposphere 1 h into their lifetimes. The cell area profiles between moderate and high MUCAPE are quite similar, suggesting the cell area is not particularly sensitive to MUCAPE once moderate values are reached. Cells occurring in more unstable environments also contain greater K_{DP} values above 4 km MSL (near 0°C level) throughout their lifetimes with the deepest K_{DP} columns in the middle of the life cycle followed by the highest values at low levels. In contrast, Z_{DR} values are greatest throughout the troposphere for moderate MUCAPE with peak values that occur earlier in the life cycle. These results suggest updrafts increase in strength as MUCAPE increases following previous studies that have shown that taller and wider Z_{DR} and K_{DP} columns (vertically oriented regions of enhanced Z_{DR} and K_{DP} above 0°C level) are correlated with the strength and size of updrafts in supercells (e.g., Kumjian et al. 2014; Snyder et al. 2017). Raindrop diameters peak within the first 30–45 min of cell lifetime like Z_{DR} values, leading the peak in rain rate that occurs 30–45 min later with the peak K_{DP} values. Interestingly, raindrop diameters are largest with moderate MUCAPE, while rain rates are much higher in the most unstable environments. These results suggest that raindrops increase in size as MUCAPE increases from low to moderate values, potentially as a result of enhanced collision–coalescence, but as the updrafts deepen and intensify with higher MUCAPE, higher rain rates are produced that may result in more collisional raindrop breakup that decreases raindrop sizes. The 90th-percentile profiles show similar correspondences with MUCAPE (not shown), suggesting the results are robust across the distribution of the cell samples.

c. Potential role of cell–cell interactions

Last, we explore the potential role of interactions between neighboring convective cells on their size and intensity. Small cells are most intense when there are more cells in their vicinity (Fig. 13a) and when there is a larger accumulated cell area in the domain (Fig. 13b). In contrast, larger cells are most intense when there are fewer cells in the surrounding area and the accumulated cell area is smaller. These results suggest that neighboring cells may promote small and weak cells to become more intense (a “collaboration” effect), but large cells may suppress development of surrounding cells (a “competition” effect).

We hypothesize that several cell-to-cell interaction processes may be producing these results. One important mechanism that leads to aggregation of deep convection is precipitation-driven cold pool interactions between nearby convective cells (e.g., Feng et al. 2015; Haerter 2019; Haerter et al. 2020; Rowe and Houze 2015; Tompkins 2001; Torri et al. 2015). Cold pool gust fronts enhance the ascent of moist and unstable boundary layer air to induce new CI (Böing et al. 2012; Khiaroudinov and Randall 2006; Rotunno et al. 1988; Wilhelmson and Chen 1982). This process may be even more effective when cold pool boundaries collide or intersect (Droegemeier and Wilhelmson 1985; Purdom 1976), which could further promote aggregation and widening of newly triggered convective updrafts (Feng et al. 2015). Greater numbers of limited size cells could be promoting cold pool interactions that support intensification of cells. On the other hand, large cells may produce strong cold pools that stabilize the environment (Del Genio et al. 2012) such that more cells result in cell weakening. In addition, more numerous limited size cells may experience less entrainment-driven dilution due to enhanced local cloudiness that moistens air surrounding updrafts, while more numerous cells may also enhance cell merging to produce wider cells less susceptible to entrainment that become more intense. However, once convection becomes sufficiently large and intense, it may suppress surrounding cells by creating strong low-level inflow and outflow that suppress inflow to other clouds and with compensating upper-level subsidence. Separating the roles of these different processes in producing cell collaboration and competition effects is a focus of future research.

6. Summary and conclusions

In this study, we characterize a wide range of deep convective cloud life cycles and their relationships with environmental conditions during the CACTI field campaign that took place between October 2018 and April 2019 in the Sierras de Córdoba (SDC) range of central Argentina. To identify and track convective cells observed in polarimetric, C-band (C-SAPR2) radar PPI volumes, we modify the FLEXTRKR algorithm (Feng et al. 2018) convective cell identification technique to (i) use composite reflectivity in complex terrain,
FIG. 12. Composite cell vertical profile evolution in (a) low-, (b) moderate-, and (c) high-MUCAPE environments. Each row shows the same variable, and the color fill represents median values. The number of cell tracks are provided in the upper-right corner in the top row. Only cells with lifetimes between 1.75 and 2.0 h that do not start from a split or end by merging are included.
tune the thresholds to break up large convective regions into separate cell components, and (iii) incorporate mean cell advection estimates between consecutive radar scans (collected at 15-min intervals) (Figs. 1 and 2). A total of 6895 cells are tracked and matched with corresponding GOES-16 satellite cloud-top property retrievals and sounding-based environmental parameters to construct a comprehensive database of cells and cell environments.

Multiple periods of prolonged convective activity with cells of varying size, intensity, and organization occurred under a wide range of environmental CAPE and moisture conditions (Fig. 4). Consistent with previous works studying terrain influence on convective cloud development, the SDC exerts a strong control on CI and growth. Convective cells preferentially initiate just east of the highest terrain during the afternoon and account for a majority of the daytime CI (Figs. 5 and 7). The number of CI events rapidly increases and peaks around midafternoon followed by a decrease in early evening, consistent with Mulholland et al. (2018). Between noon and early evening, cells gradually deepen, intensify, and expand in area (Fig. 6). The deepest cells with 30-dBZ ETHs exceeding 18 km form just east of the primary SDC ridgeline and eastward onto the plains, while the widest cells occur over the plains (Fig. 5), signifying upscale growth of cells downstream of the SDC. After sunset, CI gradually transitions to the western slope of the SDC, peaking there after midnight (Fig. 7), which is also reported by Cancelada et al. (2020) using satellite observations. This transition in CI behavior may be related to complex interactions between nocturnal strengthening of the South American low-level jet, cessation of boundary layer growth and thermal upslope flow, cold pools associated with MCSs, and the SDC topography.

Tracked cell minimum CTT is bimodal, signifying copious relatively shallow cells with cloud tops between 0º and −10ºC, more numerous deeper cells with cloud tops of −45º to −60ºC, and a fair number of cells with tops achieving −70º to −95ºC temperatures. In general, horizontal cell area, intensity (maximum reflectivity), and depth are all positively correlated, albeit substantial variability exists among these relationships (Fig. 8). The largest and most intense cells tend to occur in environments with high CAPE and low-level moisture and tend to be surface based (Fig. 9). For a given cell horizontal area, higher midtropospheric RH favors deeper cells, while for cells reaching a given depth, higher RH corresponds to smaller cells. These results suggest a drier midtroposphere may enhance entrainment-driven dilution of updraft buoyancy and limit the vertical growth of cells, consistent with theoretical and idealized modeling studies (Morrison et al. 2020; Peters et al. 2020). Greater 0–6- and 0–3-km vertical wind shear favor larger and shallower cells while weaker shear allows narrower cells to grow deeper.

Rapid cell area growth rates exhibit dependence on both large environmental wind shear and low-level moisture (Fig. 10), possibly consistent with observations of rapid cell growth in the lee of the SDC owing to terrain-induced shear enhancements (e.g., Mulholland et al. 2019, 2020; Singh et al. 2022; Trapp et al. 2020). Area growth rates were statistically similar across environments with at least modest MUCAPE, suggesting only a weak correspondence. Besides environmental factors, area growth rates also varied associated with mergers between neighboring cells or during cell splitting events, suggesting complex processes resulting in the observed cell growth rates. Both low-level relative humidity and MUCAPE strongly influence the evolution of the cell vertical structure (Figs. 11 and 12). Cells that initiate in low-level environments with more moisture are deeper during initiation, with substantially greater radar reflectivity, larger area, $Z_{DR}$, and raindrop diameter ($D_m$), indicating likely stronger initial updrafts. In moderate to high MUCAPE environments, cells grow much wider in the mid- to upper troposphere and reach higher altitudes, with larger $KDP$ and $Z_{DR}$ above the environmental freezing level, suggesting sustained development of stronger
and wider updrafts in more unstable environments. Relatively few but large raindrops occur soon after CI, followed by narrower, but more numerous raindrops and peak rainfall intensity, which is associated with a lesser \( D_m \), in high MUCAPE as compared to moderate MUCAPE conditions, which may be related to enhanced raindrop breakup in much higher rain rates at higher MUCAPE conditions. In addition, narrower cells have reflectivities that increase as the number of neighboring cells increases. In contrast, wider cells that have relatively high reflectivity have fewer neighbors (Fig. 13). We hypothesize that several processes could produce these results that deserve further research including aggregation effects from colliding cold pools, stabilization effects from stronger cold pools, reduced entrainment-driven buoyancy dilution from free-tropospheric moistening by nearby clouds and cell merging, and suppression of surrounding cells by intense convection creating strong low-level inflow and compensating upper-level subsidence.

The results presented in this study illustrate the potential use of this cell tracking database for studying interactions between convective storms and their environments. The relatively large sample size of tracked convective cells, along with their macro- and microphysical structures, and environmental conditions warrant further exploration of more detailed controls on the convective life cycle. Ongoing analyses include comprehensive examination of processes controlling CI and upscale growth, including synoptic and mesoscale circulations, aerosol properties, terrain effects, and internal convective cell precipitation structure obtained from high-resolution C-SAPR2 HSRHI scans of the tracked cells within this database. The database also demonstrates a framework to better connect measurements from various observing platforms or model analyses that facilitates comprehensive analyses needed to understand complex process interactions. This framework is currently being applied to a high-resolution regional simulation covering the CACTI field campaign (Zhang et al. 2021) to examine processes contributing to upscale growth of convection into MCSs and the results will be reported in future papers.

Acknowledgments. We thank the U.S. DOE ARM program for funding the CACTI field campaign, and all the participants of the field campaign collecting these invaluable datasets that enabled this study. We also thank three anonymous reviewers for their constructive comments and suggestions to improve the paper. This study is supported by the U.S. Department of Energy (DOE) Office of Science Biological and Environmental Research (BER) as part of the Atmospheric System Research program. Additional support was provided by National Science Foundation Grant 1661662. This research used resources of the Computer and Data Environment for Science (CADES) at the Oak Ridge National Laboratory, which is supported by the Office of Science of the U.S. Department of Energy under Contract DE-AC05-00OR22725, and the National Energy Research Scientific Computing Center (NERSC), a DOE Office of Science User Facility supported by the Office of Science of the U.S. DOE under Contract DEAC02-05CH11231. Pacific Northwest National Laboratory is operated by Battelle for the U.S. DOE under Contract DE-AC05-76RLO1830.

Data availability statement. The cell tracking database developed in this study is available through the DOE Atmospheric Radiation Measurement (ARM) Research Facility website: https://doi.org/10.5439/1844991.

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