Polarimetric Radar Metrics Related to Tornado Life Cycles and Intensity in Supercell Storms

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ABSTRACT

Polarimetric radar signatures have been related to the typical evolution of supercell storms, including through tornado life cycles. Now that polarimetric radar observations are available for a large sample of supercell storms, time series of new radar metrics can be derived. These metrics can be compared with phases of known tornado life cycles in an effort to develop new methods of anticipating tornadoes and to increase understanding of storm-scale structural and microphysical changes through supercell and tornado life cycles. In this paper, radar metrics including measures of differential reflectivity \( \textit{Z}_{\text{DR}} \) columns, \( \textit{Z}_{\text{DR}} \) arcs, polarimetrically inferred hailfall regions, and mean value of copolar correlation coefficient \( \rho_{hv} \) in the echo appendage are compared to the tornado life cycle and to storm-maximum tornado intensity in a sample of 35 tornadic supercells. It is shown that these radar metrics may change repeatedly and thus can be used to distinguish tornadic and nontornadic periods in single supercell storms, tornadogenesis from tornado demise times, and modes of storm evolution relative to tornadoes (e.g., if a storm produces one tornado or several). The polarimetric radar metrics are nearly as predictive of tornado intensity as commonly used measures of environmental variability for this sample.

1. Introduction

Numerical model solutions provide short-term guidance about potential future severe weather threats, including tornadoes (e.g., Schenkman et al. 2011; Torn and Romine 2015; Weisman et al. 2015; Sobash et al. 2016; Yussouf et al. 2016). Once convection has initiated in an environment conducive for severe weather, the primary challenge shifts to anticipation of severe weather impacts, primarily using radar observations. Traditional radar variables such as reflectivity factor at horizontal polarization \( \textit{Z}_{\text{HH}} \) and radial velocity \( \textit{V}_{r} \) have been used extensively in severe weather and tornado nowcasting. With the introduction of dual polarization to the Weather Surveillance Radar-1988 Doppler (WSR-88D) network, additional variables are now available which can be used to develop metrics informative of storm updraft and inflow characteristics. These metrics offer new possibilities for learning about ongoing storm evolution. Investigating the usefulness of polarimetric radar metrics for this purpose has value for operations and for developing a deeper understanding of storm dynamics and microphysics, including those surrounding the tornado life cycle.

Polarimetric radar features of supercell thunderstorms are generally well known and have been the focus of numerous studies (e.g., Romine et al. 2008; Kumjian and Ryzhkov 2008; Van Den Broeke et al. 2008; Kumjian et al. 2010; Snyder et al. 2013; Homeyer and Kumjian 2015; Dawson et al. 2014; French et al. 2015; and many others). These storms may produce tornadoes which, once ongoing, can be inferred by a tornadic debris signature (TDS) if one is present (e.g., Ryzhkov et al. 2005; Schultz et al. 2012; Bodine et al. 2013; Saari et al. 2014; Van Den Broeke and Jauernic 2014; Van Den Broeke 2015; and others). A radar signature informative of storm inflow characteristics is the differential reflectivity \( \textit{Z}_{\text{DR}} \) arc (e.g., Kumjian and Ryzhkov 2008; Dawson et al. 2014, 2015), which occurs along a supercell’s forward flank in a region of hydrometeor size sorting. A band of sparse large liquid drops and relatively few small drops results, with \( \textit{Z}_{\text{DR}} \) values locally exceeding 2 dB and sometimes >5–6 dB. The \( \textit{Z}_{\text{DR}} \) arc is generally observed in the lowest 2 km (e.g., Kumjian and Ryzhkov 2008, 2009). Increasing curvature of the arc and magnitude of \( \textit{Z}_{\text{DR}} \) values therein have been related to increasing tornadogenesis potential (e.g., Crowe et al.
Tornadogenesis may also occur as the $Z_{DR}$ arc is reorganizing (Palmer et al. 2011), manifest as shrinking of an existing arc, appearance of a new high-$Z_{DR}$ region along the forward flank, and its subsequent expansion (e.g., Kumjian et al. 2010). Some storms, even if weakly supercellular, may not contain a well-defined $Z_{DR}$ arc when low-level vertical vorticity is maximized (Van Den Broeke and Van Den Broeke 2015).

Features known as $Z_{DR}$ columns, marking a region of liquid drops and wet ice particles lofted by the updraft above the ambient 0°C level (e.g., Illingworth et al. 1987; Kumjian and Ryzhkov 2008; Kumjian et al. 2010, 2014; Snyder et al. 2015; and others), may change through the tornado life cycle. They may indicate updraft strengthening prior to tornadogenesis and midlevel updraft weakening and broadening leading up to tornado demise (e.g., Houser et al. 2015). A strengthening updraft may indicate increasing severe weather potential, but may not point toward the likely type of severe weather (e.g., Stano et al. 2014). Changes in the area extent of the column, and possibly its maximum altitude above the ambient 0°C level, may be used to infer changes in updraft strength (e.g., Kumjian et al. 2010; Van Den Broeke 2016, hereafter VDB16). Inferred updraft strengthening may be followed 10–15 min later by increased low-level $Z_{HH}$, possibly related to increased hailfall (e.g., Picca et al. 2010; Kumjian et al. 2014; Snyder et al. 2015).

Other polarimetric radar features have been related to tornado occurrence in supercell storms. Romine et al. (2008) found that the specific differential phase $K_{DP}$ foot, an area of high precipitation liquid water content downstream from the supercell updraft possibly representing the core of the forward-flank downdraft (Lemon and Doswell 1979), shifted downshear and expanded in areal extent leading up to the time of tornadogenesis. Van Den Broeke et al. (2008) examined the evolution of low-level polarimetric fields in a small sample of supercells through tornado life cycles, and found repeatable trends including increasing echo appendage cyclonic curvature through the tornado life cycle, and a maximum in inferred hailfall and decreased copolar correlation coefficient $\rho_{hv}$ along the supercell forward flank while a tornado was ongoing. Microphysical characteristics may vary through tornado life cycles; for instance, in a study using an X-band radar, a larger area of the appendage was dominated by large drops leading up to and during tornado dissipation (French et al. 2015). The application of polarimetric signatures to supercell storm evolution is reviewed by Kumjian (2013a).

Supercell storm evolution is largely controlled by the local environment, which may vary extensively on small spatial scales (e.g., Parker 2014). Polarimetric features have been shown to be responsive to the environment in a relatively small sample of supercell storms (e.g., VDB16). Since polarimetric radar metrics contain information about storm-scale environment and evolution, and indicate changes to updraft, inflow, and precipitation characteristics during storm cycling, it makes sense to use these observations to diagnose the tornado life cycle to the extent possible. This is especially true since some tornadoes, while closely dependent on low-level processes such as baroclinic vorticity generation, can also extend through a deep layer (e.g., Tanamachi et al. 2012). Given connections between the tornado life cycle and storm evolution noted in prior studies, it is hypothesized that polarimetric radar metrics can be used to diagnose aspects of the tornado life cycle in real time. Since prior studies have generally applied polarimetric radar features of supercell storms to the tornado life cycle using a case study approach, it is the goal of this paper to preliminarily describe relationships between tornado life cycles and the polarimetric radar metrics using a larger sample of supercell storms. Specifically, the primary findings of this study include the following:

1. Tornadic times are differentiated from nontornadic times in single storms by larger $Z_{DR}$ arcs with higher mean $Z_{DR}$ values and by smaller hail areal extent.
2. From tornadogenesis to tornado demise times the $Z_{DR}$ column areal extent increases, hail areal extent decreases, and $Z_{DR}$ arcs become larger and wider.
3. Storms that produce multiple tornadic periods have smaller and more variable hail areal extent and greater variability of the mean $Z_{DR}$ value within the $Z_{DR}$ arc.
4. Storms that produce higher-intensity tornadoes are distinguished by taller $Z_{DR}$ columns with larger areal extent and by $Z_{DR}$ arcs with higher mean pixel value and greater width variability.

These aspects of how polarimetric signatures evolve over the supercell life cycle, particularly that portion surrounding tornado production, have not been examined previously in this large a sample of storms ($n = 35$). Thus, the work presented here serves as a foundation for future work and forms a preliminary framework from which potential operational applications can be derived. Examining temporal correlations between tornadogenesis and polarimetrically inferred features such as updraft pulses and hail fallout was not a goal of this study, but may be included in future research.

2. Data and methods

Time periods were selected when at least one generally isolated, cyclonic supercell storm was present within
100 km of a polarimetric WSR-88D. Storms were identified by the presence of radar features described by Thompson et al. (2003). Since many metrics required radar data < 1 km above radar level to be calculated, preference was given to storms with lowest-elevation scan altitude < 1 km for a long time period, and at least one tornado had to occur within the analysis period. The shortest analysis period was 34 min in length, with most analysis periods exceeding 1–1.5 h in length. Events \( n = 35 \) were not chosen with regard to geographic location, and represented many regions (Fig. 1) and most months (Table 1).

Tornado reports were obtained from the Storm Events Database at the National Centers for Environmental Information (NCEI) (https://www.ncdc.noaa.gov/stormevents/). Starting and ending times of each tornado were taken as the tornadogenesis and tornado demise times, respectively, though this introduces some error. Though this database has limitations (e.g., Trapp et al. 2006), it remains the most rigorously verified tornado dataset. Level-II WSR-88D data were obtained from NCEI for each case, from the site nearest the storm of interest. Radar data were analyzed as in VDB16. Distributions of many of the radar metrics described here were not Gaussian, necessitating Wilcoxon–Mann–Whitney (WMW) statistics for comparison of metric value populations (e.g., Corder and Foreman 2014).

As in VDB16, the environment of each storm was characterized by variables (Table 2) from a representative proximity sounding, which was an initialization from the Rapid Update Cycle (RUC) or Rapid Refresh (RAP) model (as in Thompson et al. 2003, 2007). These soundings were selected to represent the undisturbed far-field environment and were from within 30 min of the center of the analysis period. Two averaged model soundings were used to obtain a representative environment if the analysis period was >1 h in length. The relatively isolated nature of many storms and a careful screening for environmental inhomogeneities seemed to preclude substantial error. Error in several RAP model output variables was estimated by Benjamin et al. (2016). For instance, the RAP dewpoint is typically 2–3°C too high, which could influence the moisture and instability measures used in this study (Table 2).

Radar data with high temporal and spatial resolution are optimal, since many supercell-associated features identified in the literature (e.g., outflow surges (Lee et al.

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**Table 1.** The date, analysis period, representative radar, and approximate location for the 35 supercell storms included in this study.

<table>
<thead>
<tr>
<th>Date of supercell</th>
<th>Analysis period (UTC)</th>
<th>Radar</th>
<th>Approximate location</th>
</tr>
</thead>
<tbody>
<tr>
<td>2 Mar 2012</td>
<td>1501–1608</td>
<td>KHTX</td>
<td>North-central Alabama</td>
</tr>
<tr>
<td>3 Mar 2012</td>
<td>0119–0233</td>
<td>KFFC</td>
<td>Northwest Georgia</td>
</tr>
<tr>
<td>14–15 Apr 2012</td>
<td>2340–0058</td>
<td>KICT</td>
<td>South-central Kansas</td>
</tr>
<tr>
<td>15 Apr 2012</td>
<td>0200–0057</td>
<td>KTWX</td>
<td>Northeast Kansas</td>
</tr>
<tr>
<td>26–27 Apr 2012</td>
<td>2337–0019</td>
<td>KOHX</td>
<td>North-central Tennessee</td>
</tr>
<tr>
<td>30 Apr 2012</td>
<td>2206–2335</td>
<td>KDDC</td>
<td>Southwest Kansas</td>
</tr>
<tr>
<td>9 Jun 2012</td>
<td>0001–0058</td>
<td>KMOT</td>
<td>Northern Michigan</td>
</tr>
<tr>
<td>18 Feb 2013</td>
<td>2253–2357</td>
<td>KSHV</td>
<td>Northeast Texas</td>
</tr>
<tr>
<td>18 Mar 2013</td>
<td>2126–2236</td>
<td>KFFC</td>
<td>West-central Georgia</td>
</tr>
<tr>
<td>31 Mar 2013</td>
<td>0137–0254</td>
<td>KINX</td>
<td>Northeast Oklahoma</td>
</tr>
<tr>
<td>15 Apr 2013</td>
<td>0020–0057</td>
<td>KTWX</td>
<td>Central Kansas</td>
</tr>
<tr>
<td>16 May 2013</td>
<td>2303–0020</td>
<td>KFWS</td>
<td>North-central Texas</td>
</tr>
<tr>
<td>19 May 2013</td>
<td>2115–2236</td>
<td>KTLX</td>
<td>Central Oklahoma</td>
</tr>
<tr>
<td>19–20 May 2013</td>
<td>2323–0001</td>
<td>KTLX</td>
<td>Central Oklahoma</td>
</tr>
<tr>
<td>14–15 Aug 2013</td>
<td>2327–0058</td>
<td>KAMA</td>
<td>Northern Texas Panhandle</td>
</tr>
<tr>
<td>28 Aug 2013</td>
<td>0316–0403</td>
<td>KDTX</td>
<td>Central Michigan</td>
</tr>
<tr>
<td>17 Nov 2013</td>
<td>1757–1830</td>
<td>KLOT</td>
<td>Northern Illinois</td>
</tr>
<tr>
<td>17 Nov 2013</td>
<td>1916–2013</td>
<td>KVWX</td>
<td>Southeast Illinois</td>
</tr>
<tr>
<td>17 Nov 2013</td>
<td>2007–2106</td>
<td>KIND</td>
<td>West-central Indiana</td>
</tr>
<tr>
<td>20 Dec 2013</td>
<td>2246–2346</td>
<td>KDGX</td>
<td>West-central Mississippi</td>
</tr>
</tbody>
</table>
TABLE 2. Variables used to characterize the environment of each supercell storm.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>MLCAPE (J kg(^{-1}))</td>
<td>Moist convective available potential energy</td>
</tr>
<tr>
<td>LCL height (m)</td>
<td>Lifted condensation level height</td>
</tr>
<tr>
<td>ESHEAR (m s(^{-1}))</td>
<td>Supercell composite parameter</td>
</tr>
<tr>
<td>3-km relative humidity (%)</td>
<td>LFC height (m)</td>
</tr>
<tr>
<td>Mean 3–6-km relative humidity (%)</td>
<td>Effective SRH (m(^2) s(^{-2}))</td>
</tr>
<tr>
<td>0°C level (m)</td>
<td>Significant tornado parameter</td>
</tr>
<tr>
<td>0–1-km SRH (m s(^{-2}))</td>
<td>Mean 6–9-km relative humidity (%)</td>
</tr>
<tr>
<td>0–1-km bulk shear (m s(^{-1}))</td>
<td>LCL temperature (°C)</td>
</tr>
<tr>
<td>Depth of layer with dewpoint depression ≥ 12°C (m)</td>
<td>Energy helicity index</td>
</tr>
<tr>
<td>Capping inversion max temperature (°C)</td>
<td>0–3-km VGP</td>
</tr>
</tbody>
</table>

2012) and precipitation cascades (Byko et al. 2009)] occur on short temporal and/or spatial scales. WSR-88D data may not resolve these features (e.g., Brotzge and Luttrell 2015). Thus, caution is required when WSR-88D data (or radar data of similar spatial and temporal resolution) are the only radar data available for use. In this study, WSR-88D data are used to provide maximum operational insight. Similar science principles apply to data collected by other platforms, though the quantitative results presented herein are best suited for use on the WSR-88D network as it is currently configured [e.g., with a 0.5° effective beamwidth characteristic of super-resolution; Brown et al. (2005)]. Smaller beamwidths are more likely to resolve localized extrema of radar variables.

Calibration of \(Z_{DR}\) is notoriously problematic for the WSR-88D network (e.g., Zrnić et al. 2006). Since deep convection was present in all datasets, scatterer-based \(Z_{DR}\) calibration was applied following Picca and Ryzhkov (2012). Regions of \(Z_{HH}\) 20–35 dBZ were sought ~1.5 km above the radar bright band. Unreliable data such as those biased by depolarization (Ryzhkov and Zrnić 2007) and differential attenuation were excluded. This \(Z_{HH}\) criterion also excludes pristine crystals, which have widely varying \(Z_{DR}\) depending on orientation. These returns are assumed to be from dry snow aggregates, with \(Z_{DR}\) averaging ~0.15 dB (Picca and Ryzhkov 2012). Values of \(Z_{DR}\) of all pixels meeting the altitude and \(Z_{HH}\) criteria in regions of good data quality were averaged, and this average value was subtracted from the expected value of 0.15 dB to obtain a calibration factor. Mean \(Z_{DR}\) bias over all radar datasets was ~0.07 dB, though the magnitude of bias exceeded 0.2 dB in 69% of events and 0.5 dB in 19% of events. \(Z_{DR}\) bias > 0.5 dB may cause some of the metrics used herein to become less reliable. Radar operators are encouraged to know their radar’s approximate \(Z_{DR}\) calibration offset prior to using this variable quantitatively. Improved calibration methods are needed and, to the extent possible, so are algorithms that are immune to \(Z_{DR}\) miscalibration.

3. Radar metrics

Quantitative radar metrics analyzed in this study are described here. Several represent areal extent values (km\(^2\)), which may be sensitive to storm size. Storm size was here defined as the base-scan area with \(Z_{HH} > 35\) dBZ, and ranged from 55 to 1962 km\(^2\) (median = 530 km\(^2\)). Areal extent measures were normalized by storm size at each analysis time. When results were compared using normalized and non-normalized values, normalization improved results for only base-scan hail areal extent. No significant difference was observed for other areal extent variables; thus, using the non-normalized magnitude of those metrics makes more sense since it is easier to measure. Polarimetric radar metrics utilized in this study are described below:

1) Normalized inferred hail areal extent (km\(^2\); hereafter hail areal extent): As illustrated by VDB16, except it has been normalized by storm areal extent and so represents percentage of storm area dominated by polarimetrically inferred hail. Hail is demarcated following prior work (e.g., Kumjian and Ryzhkov 2008; Park et al. 2009) as the region of the high-\(Z_{HH}\) (>55 dBZ) storm core with depressed \(Z_{DR}\) values (<1 dB). These thresholds were rarely modified (e.g., if \(Z_{HH}\) did not exceed 55 dBZ in the storm core, but hail was still obviously present). This metric was computed using the lowest-elevation scan if altitude of the hail region was <1 km above radar level (ARL), assuming standard beam propagation. This metric could not be computed for 4 of the 35 storms because beam altitude in the region of inferred hail exceeded 1 km.

2) Areal extent of \(Z_{DR}\) arc core (km\(^2\)): As illustrated by VDB16, the area of the \(Z_{DR}\) arc (Kumjian and Ryzhkov 2008) containing values ≥ 3.5 dB (the
“Z_{DR} arc core”) was calculated. This was done for the lowest-elevation scan, again assuming an altitude < 1 km ARL. A 3.5-dB threshold was chosen because, among these storms, it captured the core of the Z_{DR} arc while excluding surrounding high-valued pixels typically within water-coated hail and storm inflow. Temporal changes in the size of the Z_{DR} arc core were well represented by the use of this threshold. This metric could not be computed for 4 of the 35 storms because beam altitude in the Z_{DR} arc exceeded 1 km. Of the remaining 403 sample volumes, Z_{DR} arc core areal extent could not be computed for 7 (1.7%) because the forward flank moved farther away from the radar or because part of the forward flank was over the radar site.

3) Mean value of Z_{DR} arc (dB): As described by VDB16, an average value was calculated of all pixels > 0 dB within the Z_{DR} arc at the lowest-elevation scan, assuming the Z_{DR} arc was <1 km ARL. For this metric, a 2-dB threshold was used to define the Z_{DR} arc along the supercell forward flank, as pictured in VDB16’s Fig. 6b. Values of Z_{DR} should be consistently high in the size sorting region, so pixels with Z_{DR} < 0 dB likely correspond to noise and should not be included in the calculation of the mean. This metric was not computed for 3 of the 35 storms because beam altitude exceeded 1 km over part of the Z_{DR} arc. Of the remaining 414 sample volumes, this metric was not computed for 13 (3.1%) because the Z_{DR} arc was not well defined.

4) Width of Z_{DR} arc (km): As illustrated by VDB16, width of the 2-dB Z_{DR} arc was measured perpendicular to the supercell’s forward-flank Z_{HH} gradient at the lowest-elevation scan, and an average value was calculated. Altitude < 1 km ARL was required. A 2-dB threshold was chosen to be consistent with the area over which the mean Z_{DR} value was calculated (metric 3). This metric was not computed for 5 of the 35 storms because of radar beam altitude. Of the remaining 392 sample volumes, this metric was not computed for 6 (1.5%) because the 2-dB Z_{DR} arc was not fully <1-km altitude.

5) Areal extent of Z_{DR} column (km²): As illustrated by VDB16, the elevation closest to 1 km above the ambient 0°C level [determined using RUC/RAP sounding(s)] was determined, and the area of the 0.5-dB Z_{DR} column there was calculated by summing the area of all pixels reaching or exceeding this threshold in the column. This threshold is not the same as the 1-dB threshold used in Snyder et al. (2015) for Z_{DR} column maximum altitude. This is because, at an altitude ~1 km above the ambient 0°C level, most hydrometeors surrounding the updraft are likely dry snow and therefore typically have Z_{DR} values of 0–0.5 dB. Areas with Z_{DR} values > 0.5 dB in this situation are likely associated with liquid, and therefore provide a reasonable demarcation of the updraft region. A 1-dB threshold would have eliminated some or much of the updraft region in some storms. Various threshold values should, however, be tested in future work. This metric was not computed for one storm because noisy Z_{DR} aloft made it difficult to determine the boundaries of the 0.5-dB Z_{DR} column. Of the other 496 sample volumes, this metric was not computed for 31 (6.25%) because of similar concerns about the ability to accurately demarcate the 0.5-dB Z_{DR} column.

6) Altitude > 0°C of Z_{DR} column (km): As illustrated by VDB16, the altitude at the top of the 1-dB Z_{DR} column was identified, and the altitude of the ambient 0°C level was subtracted from this value. A 1-dB threshold is consistent with the Z_{DR} column algorithm presented by Snyder et al. (2015). Though different threshold values could be chosen that may yield different results (and should be tested in future work), keeping consistency with the algorithm of Snyder et al. (2015) is deemed important so future work, which is likely to use that algorithm, is comparable. This method assumes that the Z_{DR} column top occurs at beam centerline of the highest tilt at which the column appears, which can be subject to substantial error given vertical beam spreading, especially at a large range. Values of this metric were not calculated for 24 of the 517 sample volumes (4.6%) because of an inability to clearly identify the top of the 1-dB column.

7) Mean lowest-elevation scan appendage ρ_{hv} value: Demarcating the appendage as in French et al. (2015), the mean ρ_{hv} value was calculated. This metric was not calculated if contamination by debris, three-body scatter (Kumjian 2013b), or nonuniform beam filling (e.g., Ryzhkov 2007; Kumjian 2013b) was suspected. The noisy nature of ρ_{hv} limited the usefulness of this metric in some cases. In total, it was calculated for 274 sample volumes from 27 storms (53% of available volumes).

8) Maximum storm core Z_{HH} value (dBZ): The highest Z_{HH} value within the storm core was recorded if altitude was <1 km. This simple metric is thought to provide information on the intensity of precipitation, including hail, within the storm core. Fluctuations of precipitation intensity may be related to a storm’s ability to generate outflow. This variable may be prone to error due to noise and uncertainty in what scatterers contribute to the maximum value. This
metric was not recorded for 4 of the 35 storms because the radar beam was too high.

As in VDB16, the coefficient of variation was utilized as a measure of variability of the radar metrics and was calculated for all the metrics described above. For a set of values (e.g., all inferred hail areal extent values for a given storm), the coefficient of variation is the standard deviation of the set of values divided by the set’s mean value. Normalizing by the mean allows a measure of variability to be compared between storms with dissimilar mean values. Though not an issue with the metrics examined here, any linear temporal trends should also be removed from time series of radar metrics prior to computing variability (e.g., decreasing $\rho_{hv}$ as a storm moves farther from the radar site).

4. Supercell environments

The storms analyzed occurred in a variety of environments, generally spanning the parameter space expected of tornadic supercell storms (e.g., Rasmussen and Blanchard 1998; Rasmussen 2003; Thompson et al. 2003). Knowing the range of storm environments represented is relevant to the goals of this study primarily because an attempt is made to show how the radar metrics are related to tornado characteristics, and such application would ideally be useful over a large range of supercell environments.

Figure 2 illustrates how some commonly used environmental parameters varied across the 35 storms analyzed. Most unstable convective available potential energy (MUCAPE; Evans and Doswell 2001; Fig. 2a) varied from $2\ J\ kg^{-1}$ to near $4951\ J\ kg^{-1}$, with values for most storms $\sim 650$ to $2000\ J\ kg^{-1}$. This range covers the spectrum of supercell environments, except for uncommon environments with MUCAPE $> 5000\ J\ kg^{-1}$. The sounding with only $2J\ kg^{-1}$ of MUCAPE may be unrepresentative of the immediate storm environment. Height of the level of free convection (LFC) typically ranged from just above 500 m to just above 3500 m (Fig. 2b), also representing most supercell environments (e.g., Davies 2004). One extreme-LFC case (4595 m) was associated with a short-lived April EF0 tornado in Kansas. Storm-relative helicity (SRH) in the 0–1-km bulk shear (ESHEAR; Thompson et al. 2007; Fig. 2d) ranged from 0 to near $60\ m^2 s^{-2}$, as expected for tornadic supercell environments (e.g., Rasmussen 2003). Effective bulk shear (ESHEAR; Thompson et al. 2007; Fig. 2d) ranged up to $25\ m s^{-1}$, again in accord with prior observations from tornadic supercell environments. Finally, mixed-layer convective available convective energy (MLCAPE; Thompson et al. 2003) was plotted against 0–6-km shear for each environment (Fig. 2e) to be comparable to supercell environments examined by Rasmussen and Blanchard (1998). Much of the parameter space for supercell storms and tornadic storms found by Rasmussen and Blanchard (1998) was also represented by storms in this study, with the exception that tornadic storms in this study tended to be associated with larger mean 0–6-km shear.

5. Polarimetric radar metrics and tornado characteristics in supercell storms

a. Comparison of metric values to prior study

Several metrics used here were introduced by VDB16 for a smaller sample of supercell storms. To show that values of the metrics are statistically similar between the population of storms used in VDB16 (15 storms) and the storms used in this study (35 storms), comparisons are made between populations of the metric values. WMW $p$ values were computed (Table 3) using each study as a separate population. All $p$ values were $>0.30$, showing similar populations of metric values between the studies, except for $Z_{DR}$ column maximum altitude above the ambient 0°C level ($p = 0.041$; Table 3). The value of this metric averaged 2.29 km for storms in VDB16, and 2.77 km in the present study. Since storms averaged approximately the same distance from the radar in each study, this difference is unlikely to be attributable to differing magnitude of error in estimates of this variable at different ranges. A larger percentage of storms in the present study had $Z_{DR}$ columns $> 3$ km above the ambient 0°C level. Otherwise, no systematic differences were present between the two studies for these radar metrics.

b. Analysis across the tornado life cycle

Radar sweeps for each storm were categorized into tornadic and nontornadic, where “tornadic” meant that the Storm Events Database indicated a tornado was ongoing. “Nontornadic times” refers to sample volumes from a tornadic storm when a tornado was not ongoing. Tornadic analysis periods were hypothesized to be distinguishable by having larger $Z_{DR}$ arcs with higher mean $Z_{DR}$ values, and by larger hail areal extent. Of the 35 storms, 16 had at least 3 analysis times in each category and were included in this analysis. Mean values of the radar metrics were computed for tornadic and nontornadic analysis periods. The same computations were made using only storms that produced EF2+ tornadoes ($\sim 8$ storms could be included for most metrics), to see if storms with stronger tornadoes showed stronger trends in the radar variables. This did not substantially change
FIG. 2. Range of environments among storms analyzed: (a) MUCAPE (J kg⁻¹), (b) LFC height (m), (c) 0–1-km SRH (m² s⁻²), and (d) ESHEAR (m s⁻¹). The bottom of the blue box in each panel is the first quartile, and the top of the blue box is the third quartile. The orange bar indicates the median value, while the red plus sign marks the mean value. Bars are at the 9th and 91st percentiles, with outliers indicated as circles even farther removed from the median. (e) A scatterplot of MLCAPE (J kg⁻¹) vs 0-6-km shear (m s⁻¹), as in Rasmussen and Blanchard (1998).
the results, so findings here are for all storms. The relatively small number of storms means that the results presented here should be considered preliminary, but they yield information that should be investigated further.

As hypothesized, areal extent of the 3.5-dB $Z_{DR}$ arc core and mean $Z_{DR}$ value within the arc were larger in at least two-thirds of storms during tornadic times (Fig. 3), consistent with prior findings (Palmer et al. 2011; Crowe et al. 2012). Mean difference in $Z_{DR}$ arc core areal extent was $\sim 16$ km$^2$, corresponding to $\sim 99$ radar pixels at a range of 60 km. One-third of storms had a WMW $p$ value $< 0.15$ for this metric when tornadic times were compared to nontornadic times (indicating that tornadic and nontornadic times were statistically distinguishable). Normalized hail areal extent was $\sim 12$ km$^2$ larger at nontornadic times on average (Fig. 3), corresponding to $\sim 74$ pixels at a range of 60 km. Relatively few storms, however, had a WMW $p$ value $< 0.15$ for this metric. The storm-core maximum $Z_{HH}$ value was larger at nontornadic times in approximately three-fourths of storms (Fig. 3), consistent with larger hail areal extent. The $Z_{DR}$ column metrics did not show repeatable differences between tornadic and nontornadic times for this sample of storms. Finally, though mean $Z_{DR}$ arc width was only $\sim 0.5$ km larger for tornadic times, 56% of storms had a statistically meaningful difference between tornadic and nontornadic analysis periods (WMW $p$ value $< 0.15$; Fig. 3).

Trends were also sought in the radar metrics across tornado life cycles, as such trends may increase predictability and may provide insight on storm-scale processes surrounding the tornado life cycle. Moving from tornadogenesis to tornado demise, it was hypothesized that $Z_{DR}$ column maximum altitude should increase, hail areal extent should decrease, and $Z_{DR}$ arcs should become smaller with lower-magnitude mean values. Tornadogenesis times were defined as being from 4 min prior to 4 min after reported genesis, while tornado demise times were defined as being from 4 min prior to 4 min after reported demise. A 4-min threshold was used to reflect the approximate radar update time of most events. A longer threshold would have resulted in the inclusion of less-representative times in the analysis, while a shorter threshold would have limited the number of sample volumes. If a radar sweep met the definition for both a genesis and demise time, it was assigned solely to the category to which it was temporally closest. Storms were retained ($n = 30$) if they had at least two analysis periods in each of at least two tornado life cycle categories. Mean values of the radar metrics in each category are provided in Table 4.

Maximum $Z_{DR}$ column altitude above the ambient 0°C level decreased from genesis to demise times in 67% of storms (Table 4), contrary to the hypothesized pattern. Areal extent of the 0.5-dB $Z_{DR}$ column $\sim 1$ km above the ambient 0°C level averaged $\sim 8$ km$^2$ less at genesis times (Table 4), corresponding to $\sim 49$ pixels at a range of 60 km, and consistent with the hypothesis that convective updrafts may weaken around tornadogenesis (e.g., Adler and Fenn 1981; Dowell and Bluestein 1997; Picca et al. 2015). Findings for both metrics indicated small enough differences that operational implementation will likely not be possible. Normalized hail extent was repeatedly highest during tornadogenesis times and decreased through tornado demise (Table 4), as hypothesized and similar to the findings of Van Den Broeke et al. (2008). The difference between genesis and demise times was weakly significant ($p = 0.083$), and the trend in normalized hail extent may be operationally useful. Areal extent of the $Z_{DR}$ arc core increased through the tornado life cycle (Table 4), and for this sample of storms averaged $\sim 14$ km$^2$ lower at genesis times, corresponding to $\sim 87$ pixels at a range of 60 km. This difference was large enough to possibly have operational value in some cases, but was not statistically significant. Similarly, $Z_{DR}$ arc width increased through the tornado life cycle (Table 4), averaging 0.7 km wider at demise times than at genesis times. This difference may be visible in some real-time situations.

c. Analysis by mode of tornado production

While some supercell storms have discrete tornadic periods, others produce tornadoes nearly continually. Operationally, there may be value in having an indication that a storm may exhibit temporal gaps between tornadoes, regardless of tornado intensity. It is
hypothesized that storms with multiple tornadic periods will have larger, taller updrafts, and larger variability of the radar metrics. In this study, storms are classified into two modes depending on whether there are temporal gaps between tornadoes. This differs from the modes defined by Tanamachi et al. (2012), where mode 1 (characterized by EF3 or weaker tornadoes produced nearly regularly) and mode 2 (characterized by the production of at least one EF4 or stronger tornado). In this study, the two modes are defined as follows:

Mode A: From the time of initial tornadogenesis to final tornado demise, a tornado is always ongoing. This tornadic time period may contain one or more tornadoes of any intensity, but multiple tornadoes must temporally overlap. A total of 23 storms fit this definition.

Mode B: From the time of initial tornadogenesis to the time of final tornado demise, there are one or more time periods when a tornado is not ongoing. A tornadic period may contain one or more tornadoes of any intensity. The remaining 12 storms fit this definition.

Populations of averages of the radar metrics for each mode were compared using WMW \( p \) values (summarized in Fig. 4). The hypothesized increase in updraft height and areal extent was generally not realized among these storms. Maximum \( Z_{\text{DR}} \) column altitude above the 0°C level and maximum \( Z_{\text{HH}} \) value within the storm core seemed to somewhat differentiate the two storm categories (Fig. 4a), though \( p \) was relatively high for these metrics (0.10 < \( p \) ≤ 0.15). The metric most strongly differentiating mode A and mode B storms was normalized hail areal extent, which was smaller and much more variable in mode B storms (Fig. 4b), and mode B storms contained slightly lower mean \( Z_{\text{DR}} \) values within the arc. This may result from the generally more extensive \( Z_{\text{DR}} \) arc in tornadic storms. Note that the

![Fig. 3. Average metric values during times when a tornado was ongoing (red dots) and when a tornado was not known to be ongoing (blue dots). Values for 3 of the metrics were divided by 10 to fit on the same axes, indicated by the green “+[10].” Normalized hail areal extent was multiplied by 100 to fit on the axis, indicated by the green “[×100]” (e.g., a value of 4.0 indicates that 4.0% of the storm area was inferred to be hail at base scan). Yellow stars indicate metrics for which the observed pattern (tornadic times larger or smaller) was present in at least two-thirds of cases (percentage indicated within star; value ≥ 67% indicates tornadic times larger). Percentage of storms for which the WMW \( p \) value comparing tornadic and nontornadic times was <0.15 is indicated below the name of each metric.](image.png)
difference in mean \( Z_{DR} \) between mode A and B storms was \(<0.1 \text{ dB} \), so this finding will be difficult to use operationally.

d. Analysis by tornado intensity

Another hypothesis of this study relates the radar metrics to maximum tornado intensity in supercell storms. Benefit may be realized if radar signatures can be used to differentiate storms that produce relatively strong versus weak tornadoes. Following is a description of how the radar metrics vary by maximum intensity category of tornadoes in each storm, including a description of what signatures distinguish weak-tornado storms and significant-tornado storms. Tornado intensity is also related to storm environments.

Initially, the set of 35 storms was divided into those that produced at least one “significant” tornado (\( \geq \text{EF2; } n = 13 \)), and those that produced less intense tornadoes (\( \text{EF0/1; } n = 22 \)). The radar metrics generally did not distinguish well between these storm categories, with a few exceptions. Maximum altitude of the \( Z_{DR} \) column above the ambient 0°C level most strongly distinguished these categories and was higher in significant-tornado storms (2.55 km vs 3.16 km; \( p = 0.022 \)); this difference did not appear due to a difference in radar-storm distance, which was statistically similar between these categories).

Areal extent of the \( Z_{DR} \) column ~1 km above the ambient 0°C level was larger (47.9 km² vs 67.5 km²; \( p = 0.148 \); corresponding to ~124 pixels at a range of 60 km) and less variable (\( \text{variability} = 0.41 \text{ vs 0.31; } p = 0.072 \)) in storms that produced significant tornadoes. Updrafts among this sample of significant-tornado storms were taller and larger, which seemed to be the best way to distinguish them from other tornado-producing storms using these radar metrics. Mean value of \( Z_{DR} \) within the \( Z_{DR} \) arc was slightly higher (2.61 dB vs 2.86 dB; \( p = 0.182 \)) and more variable (\( \text{variability} = 0.09 \text{ vs 0.11; } p = 0.182 \)) in storms that produced significant tornadoes. The \( Z_{DR} \) arc width variability was greater in significant-tornado storms (0.21 vs 0.27; \( p = 0.064 \)). These results indicate that storms that produce significant tornadoes exhibit greater variability of many radar metrics, but lesser variability of metrics related to updraft intensity (\( Z_{DR} \) column extent and altitude). Hail extent did not distinguish these storm categories, contrary to the hypothesized pattern.

Next, the same hypothesis was tested by examining whether weak-tornado storms (maximum intensity rating EF0; \( n = 15 \)) and significant-tornado storms (intensity rating EF3+; \( n = 7 \)) are differentiable from other tornado-producing storms. The populations of radar metrics from storms in these categories were compared to the corresponding populations from all other storms in the dataset using WMW statistics. Mean metric values and comparison \( p \) values are shown in Fig. 5. Storms producing EF0 tornadoes were distinguished from other tornadic storms by having \( Z_{DR} \) columns that did not extend as high above the ambient 0°C level, lower mean \( Z_{DR} \) values in the \( Z_{DR} \) arc, less variable maximum \( Z_{HH} \) values in the storm core, and less variable mean \( p_{hv} \) in the echo appendage (Fig. 5). These findings support the hypothesis of weaker updrafts and inflow, and generally less variable storm processes, in weakly tornadic storms.

Storms that produced EF3+ tornadoes exhibited much larger \( Z_{DR} \) column areal extent at ~1 km above the ambient 0°C level, more variable mean values of \( Z_{DR} \) arc width and \( Z_{DR} \) arc mean pixel value, and higher mean \( Z_{DR} \) values within the \( Z_{DR} \) arc (Fig. 5), as hypothesized. Magnitude of \( Z_{DR} \) arc values was not statistically distinguishable between strong-tornado and other storms if uncalibrated \( Z_{DR} \) values were used, underscoring the criticality of accounting for the \( Z_{DR} \) calibration offset. Variability of \( Z_{DR} \) column areal extent decreased with tornado intensity, indicating consistently strong updrafts in storms that produced significant tornadoes (also evidenced by the large \( Z_{DR} \) column areal extent in storms that produced significant tornadoes; Fig. 5a). Finally, \( Z_{DR} \) arc width increased with tornado intensity (Fig. 5a), possibly indicating stronger storm-relative winds in the inflow layer of these storms. The difference in mean \( Z_{DR} \) arc width was \( >1 \text{ km} \) between storms producing EF0 and EF3+ tornadoes, a large enough difference to possibly be operationally useful in some circumstances.

<table>
<thead>
<tr>
<th>Radar metric</th>
<th>Genesis times</th>
<th>Tornado ongoing</th>
<th>Demise times</th>
</tr>
</thead>
<tbody>
<tr>
<td>Normalized hail areal extent</td>
<td>0.052</td>
<td>0.043</td>
<td>0.031</td>
</tr>
<tr>
<td>( Z_{DR} ) arc core extent (km²)</td>
<td>72.39</td>
<td>82.57</td>
<td>86.10</td>
</tr>
<tr>
<td>( Z_{DR} ) arc mean pixel value (dB)</td>
<td>2.86</td>
<td>2.91</td>
<td>2.85</td>
</tr>
<tr>
<td>( Z_{DR} ) arc width (km)</td>
<td>7.70</td>
<td>7.79</td>
<td>8.40</td>
</tr>
<tr>
<td>( Z_{DR} ) column extent (km²)</td>
<td>55.90</td>
<td>57.72</td>
<td>63.70</td>
</tr>
<tr>
<td>( Z_{DR} ) column max altitude &gt; 0°C (km)</td>
<td>3.08</td>
<td>2.84</td>
<td>2.90</td>
</tr>
<tr>
<td>Storm core ( Z_{HH} ) max value (dBZ)</td>
<td>63.93</td>
<td>63.08</td>
<td>62.24</td>
</tr>
</tbody>
</table>
To assess the ability of the radar metrics to distinguish maximum tornado intensity relative to knowledge of environmental variables, the storm-scale environment was compared for storms in the three tornado intensity categories. A sample of 29 environmental variables was included (Table 2); WMW $p$ values were calculated for storms that produced EF3+ tornadoes versus those that produced weaker tornadoes. The most significant environmental variables for distinguishing significant-tornado storms, in this sample, were the energy helicity index (Rasmussen 2003; $p = 0.036$), the supercell composite parameter (Rasmussen 2003; $p = 0.049$),
Fig. 5. As in Fig. 4, but for tornado intensity categories. Blue dots are values for EF0 tornadoes, yellow dots are values for EF1/2 tornadoes, and red dots are values for EF3+ tornadoes. The first WMW $p$ value below each metric treats EF0 tornadoes as a separate population, while the second WMW $p$ value treats EF3+ tornadoes as a separate population. Italicized bold $p$ values highlight values $< 0.10$. 

$\text{a) Hail Areal Extent (km}^2 \geq 10\text{)}$ $\text{b) Hail Areal Extent (km}^2\text{)}$

<table>
<thead>
<tr>
<th>Metric</th>
<th>EF0</th>
<th>EF1/2</th>
<th>EF3+</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hail Areal Extent (km$^2$)</td>
<td>0.828</td>
<td>0.795</td>
<td>0.548</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Areal Extent (km$^2$)</td>
<td>0.984</td>
<td>0.238</td>
<td>0.942</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Mean Pixel Value (dB)</td>
<td>0.126</td>
<td>0.305</td>
<td>0.072</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Column Areal Extent (km$^2$)</td>
<td>0.787</td>
<td>0.305</td>
<td>0.292</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Column Width (km)</td>
<td>0.204</td>
<td>0.231</td>
<td>0.039</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Storm Core Maximum Value (dBZ)</td>
<td>0.312</td>
<td>0.051</td>
<td>0.159</td>
</tr>
<tr>
<td>$Z_{\text{Arc}}$ Column Maximum Extent &gt; 0°C (km)</td>
<td>0.064</td>
<td>0.564</td>
<td>0.300</td>
</tr>
<tr>
<td>Mean Appendage $\rho_v$ Value ($^\circ$ 2)</td>
<td>0.460</td>
<td>0.083</td>
<td>0.951</td>
</tr>
<tr>
<td>Mean Appendage $Z_{\text{Arc}}$ Value (dB)</td>
<td>0.676</td>
<td>0.336</td>
<td>NA</td>
</tr>
</tbody>
</table>
MLCAPE (Rasmussen and Blanchard 1998; \( p = 0.067 \)), 0–6-km shear (Rasmussen and Blanchard 1998; \( p = 0.081 \)), and the vorticity generation parameter (VGP) calculated using the 0–3-km layer (Rasmussen and Blanchard 1998; \( p = 0.086 \)). All other environmental variables had \( p > 0.10 \). Compared to the radar metrics, knowledge of the storm-scale environment may allow a slightly greater ability to distinguish significant-tornado from other tornadic storms among this sample. The best environmental variable (energy helicity index, \( p = 0.036 \)) performed similarly to the best radar metric (\( Z_{\text{DR}} \) column areal extent, \( p = 0.039 \)). If other radar metrics were incorporated (e.g., using \( K_{\text{DP}} \) information), it seems possible that radar metrics could be as predictive of storm-maximum tornado intensity as knowledge of environmental variables.

6. Summary and discussion

Quantitative radar metrics were examined over a sample of tornadic supercell storms, including tornado life cycles and intensity. The metrics can be readily estimated using WSR-88D data and are mostly related to storm updraft and inflow characteristics. Though metrics that are measures of areal extent may be sensitive to storm size given the wide variation in observed supercell morphology, normalization by storm size did not meaningfully influence results except for hail areal extent. Thus, given the results for this sample of storms it is recommended to use nonnormalized values for \( Z_{\text{DR}} \) arc areal extent and \( Z_{\text{DR}} \) column areal extent aloft, but to normalize hail areal extent by the 35-dBZ storm size. Future studies should check whether normalization is helpful for radar metrics that are measures of areal extent and should explore in more detail the value for tornado predictability of other metrics such as those related to storm shape, \( K_{\text{DP}} \), and variations in \( Z_{\text{DR}} \) across the appendage through tornado life cycles. Mean \( Z_{\text{DR}} \) value in the appendage region was calculated for each analysis time in this study, and no significant results were found (not shown). A \( Z_{\text{DR}} \)-\( Z_{\text{HH}} \) pairing approach such as that of Kumjian (2011) and French et al. (2015) may be more insightful.

Relationships between inferred hail areal extent and \( Z_{\text{DR}} \) arc areal extent/mean pixel value were examined, since disruptions of the \( Z_{\text{DR}} \) arc have been linked to hail fallout in past studies (e.g., Kumjian and Ryzhkov 2009; Palmer et al. 2011; Picca and Ryzhkov 2012; Tanamachi and Heinselman 2016). Average values of Pearson’s correlation coefficient between series of the appropriate variables were low (for hail extent vs \( Z_{\text{DR}} \) arc areal extent, \( n = 28 \) and \( r = -0.10 \); for hail extent versus mean \( Z_{\text{DR}} \) arc pixel value, \( n = 29 \) and \( r = 0.03 \)). While hail descent was closely related to \( Z_{\text{DR}} \) arc disruption in some storms (not shown), this was not repeatable across the sample of storms. As noted by Tanamachi and Heinselman (2016), identification of the hail region was often facilitated by \( Z_{\text{DR}} \) contrast with relatively higher values in the adjacent \( Z_{\text{DR}} \) arc, since these two features were often contiguous.

For individual storms, tornadic analysis times were distinguishable from nontornadic analysis times by smaller hail areal extent, a larger 3.5-dB \( Z_{\text{DR}} \) arc and a larger mean \( Z_{\text{DR}} \) value within the arc, and smaller storm-core maximum \( Z_{\text{HH}} \) value. These findings generally support the hypothesis of more robust storm-relative inflow when a tornado is ongoing, consistent with prior research (Palmer et al. 2011; Crowe et al. 2012). The hypothesis that hail extent would be larger during tornadic times (e.g., Van Den Broeke et al. 2008) was not supported, though the finding of smaller hail extent during tornadic times was not statistically significant (\( p = 0.30 \)). Weak \( Z_{\text{DR}} \) column differences between tornadic and nontornadic times echo the results of Picca et al. (2015), and may reflect difficulty in estimating \( Z_{\text{DR}} \) column metrics given beam spreading with height. Some scanning strategies (e.g., volume coverage pattern 212) also have coarse vertical midlevel sampling at large range (Snyder et al. 2015).

The tornado life cycle was weakly distinguished by radar metrics among this sample of storms, and most metrics did not display differences that were likely to be operationally useful. Hypothesized updraft weakening around tornadogenesis was not supported, as \( Z_{\text{DR}} \) column maximum altitude slightly decreased toward demise times (though the decrease was not statistically significant). Prior results in the literature have indicated slight weakening, and given the coarse vertical \( Z_{\text{DR}} \) resolution, changes in updraft altitude through tornado life cycles may not be reliable. The \( Z_{\text{DR}} \) arc mean pixel value and width increased through the tornado life cycle, contrary to the hypothesized pattern of a larger and more intense \( Z_{\text{DR}} \) arc around tornadogenesis (e.g., Palmer et al. 2011; Crowe et al. 2012), though these findings also were not statistically significant. Normalized hail extent was largest at tornadogenesis times, decreasing in a statistically significant way through tornado demise.

Two storm modes were defined: in mode A supercells, there is only one tornadic period, while in mode B supercells, there are several periods in which a tornado or tornadoes occur. Measures of metric variability were often larger in mode B storms, as hypothesized. Hail areal extent was smaller in mode B storms, which was not hypothesized, and no speculation is presented as to why this may be the case. The findings did not strongly support the hypothesis that updrafts would be larger and
taller in mode B storms, though $Z_{DR}$ column maximum altitude above 0°C was higher in mode B storms ($p = 0.151$; Fig. 4).

The ability of radar metrics to provide guidance on tornado intensity was investigated. Significant-tornado storms (EF2+) were distinguished by large $Z_{DR}$ arc width variability and $Z_{DR}$ column maximum altitude above the ambient 0°C level. As hypothesized, large and tall updrafts distinguished significant-tornado storms reasonably well among this sample of storms. Storms producing weak (EF0 only) tornadoes, as hypothesized, were distinguished by generally small measures of variability and by $Z_{DR}$ columns extending to lower maximum altitude. Conversely, storms producing EF3+ tornadoes were distinguished by large measures of variability, including that of $Z_{DR}$ arc width and mean $Z_{DR}$ value in the arc. Areal extent of the $Z_{DR}$ column ~1 km above the ambient 0°C level was >70% larger in significant-tornado storms than in other tornadic storms examined, and mean $Z_{DR}$ value within the $Z_{DR}$ arc was ~9% (0.25 dB) larger. Storms that produce significant tornadoes, as hypothesized, were associated with large, strong updrafts and large-magnitude temporal changes to values of the polarimetric radar metrics. Hail extent variability was hypothesized to be larger in significant-tornado storms, but this was not supported by the findings from this sample of storms. Preliminary work here suggests that tornado intensity may be as predictable using the radar metrics as using knowledge of environmental conditions, though both should optimally be utilized together. Tornadic and nontornadic supercells may occur in sufficiently close proximity that the observational network is not useful to differentiate their environments (e.g., Klees et al. 2016). Severe weather predictability can be very sensitive to mesoscale details (e.g., Bluestein and Snyder 2015), and thus any attempt to use radar information for this purpose must be supplemented by other observational data and model guidance.

While many results presented here were statistically significant, most comparisons examined were not. This can be explained by the large variability inherent to values of the radar metrics. Thus, there was often substantial overlap between storms in two categories being compared, resulting in weaker statistical significance. This natural variability may be a key limitation to using radar metrics to predict tornado characteristics of particular storms, but the magnitude of this variability may also be a source of useful information. In the future, as more data become available and automated algorithms are developed, it may become possible to develop a model to predict tornadogenesis and tornado demise times given trends in these and other radar metrics. Future work should also explore microphysical processes and differences leading to the observed results. Finally, temporal correlations should be explored between tornadogenesis and radar-inferred events such as updraft pulses and bursts of hail reaching the surface.

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