

UNIVERSITY OF OKLAHOMA  
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HIGH-TEMPORAL RESOLUTION X-BAND POLARIMETRIC RADAR  
ANALYSIS OF THE 20 MAY 2013 MOORE, OKLAHOMA SUPERCELL  
DURING TORNADOGENESIS AND TORNADO INTENSIFICATION

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A THESIS APPROVED FOR THE  
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# Abstract

On 20 May 2013, a long-track violent EF-5 tornado impacted Moore, Oklahoma and surrounding areas. Due to the relatively high frequency of severe weather in this urban region, there exists an extensive network of radars that were operating in close proximity to the supercell throughout its lifecycle. This network of radars consists of mainly experimental and operational NEXRAD WSR-88Ds: 1) Twin Lakes, Oklahoma (KTLX), 2) Norman, Oklahoma (KOUN), and 3) Norman, Oklahoma (KCRI). Additionally, PX-1000, a transportable, X-band, dual-polarimetric radar, was operating at the University of Oklahoma Westheimer Airport in a 2.6 single-elevation PPI scanning strategy, providing updates every 20 s from 19:14:59 to 20:39:51 UTC. While previous studies have focused on the mature period of the tornadic supercell (Burgess et al. 2014; Kurdzo et al. 2015), there has been a lack of studies detailing the early evolution of the supercell.

The objective of this study is to analyze the dual-polarization signatures to detail storm-scale processes that occurred over short time periods between (i) the early evolution of the supercell to tornadogenesis, 19:30 UTC to 19:56 UTC, and (ii) from tornadogenesis through tornado intensification, 19:56 UTC to 20:08 UTC. These processes are extracted using the high-spatiotemporal resolution data from the PX-1000 to examine low-level processes during tornadogenesis and intensification. A comparative analysis using contemporaneous observations from the nearby WSR-88Ds, and PX-1000 are presented. The evolution of the ZDR arc and KDP

foot structure, vertical extent of the ZDR and KDP columns, tornado debris signature characteristics, radial velocity signatures, and hydrometeor classification are investigated as evidence of precursors to tornadogenesis and the subsequent rapid intensification.

Key findings include the presence of erratic tornado debris signatures observed as early as 15 minutes prior to tornadogenesis. Radial velocity analyses are presented to determine if low-level rotation was sufficient to meet tornado criteria during this period and examine the transient behavior of this signature. Trends in the vertical extent of ZDR and KDP columns are also analyzed to infer the presence and robustness of mid-level updrafts. A ZDR column split was evident in the same period as the erratic tornado debris signatures, further suggesting that there may have been transient intensifications in shallow layers prior to tornadogenesis. Additionally, trends in ZDR arc and KDP foot spatial structure, magnitudes of peak values, and the separation of these signatures are analyzed for the periods before and after tornadogenesis. Hypotheses of the influence of a storm merger as well as hail versus debris fallout are also discussed.

# Chapter 1

## Introduction

Radars have been a prominent means of remote sensing of the atmosphere since as early as the 1940s. In 1988, the Weather Surveillance Radar - 1988 Doppler (WSR-88D) radar network was established as a national standard for operational use. The applications of radars have increased as new algorithms are developed and technology advances (e.g., Seliga and Bringi 1976; Herzegh and Jameson 1992; Bringi and Chandrasekar 2001; Moisseev and Chandrasekar 2007; Thompson et al. 2018). The recent upgrade to dual-polarization capabilities has been the latest of such advancements. Further advancements have included the development and use of mobile radars in the field.

Longer wavelength radars are not subject to attenuation, but require a much larger antenna and therefore are more expensive and not mobile. A prominent example of longer wavelength, S-band radars are the WSR-88D radar network which operate at a 10-cm wavelength. Shorter wavelengths, such as the 3-cm X-band wavelength, are subject to attenuation. However, the antenna can be small enough to make it ideal for transport and the radar can provide higher resolution data. Smaller, lighter antennas also mean that the radar can acquire a higher rotation rate and therefore, this wavelength is suitable for mobile research radars such as RaXPoL and the PX-1000 (Pazmany et al. 2013; Kurdzo et al. 2015).

The objective of this present study is to analyze the dual-polarization signatures to detail storm-scale processes that occurred over short time periods between (i) the early evolution of the supercell to tornadogenesis, 19:30 UTC to 19:56 UTC, and (ii) from tornadogenesis through tornado intensification, 19:56 UTC to 20:08 UTC. This study utilizes data collected by the PX-1000 and an adjacent network of WSR-88Ds that provides a comparative analysis spatiotemporally. The analysis presented is organized chronologically to better explain the evolution of the supercell. Chapter 2 provides an overview of dual-polarimetric radars and will be supplemented with a discussion of past studies, especially in context of supercells and tornadoes. Chapter 3 discusses more about the data and methods used for the analysis. Chapter 4 presents the analysis. Chapter 5 presents conclusions with recommendations on future work.

## Chapter 2

### Background

The pulse repetition time (PRT) is the amount of time between two subsequent pulses and is typically on the order of about 1 ms. Because EM pulses travel at the speed of light, the maximum unambiguous range is given by

$$r_a = \frac{cT_s}{2} \quad (2.1)$$

where  $T_s$  is the pulse repetition frequency and  $c = 3 \times 10^8 \text{ m s}^{-1}$  is the speed of light. The factor of 2 in the denominator is necessary to take into account that the beam has to travel to the radar beam and back. It is possible that scatterers beyond this maximum unambiguous range create echoes detectable by the radar, but the radar will have sent out another pulse by this time. Therefore, the radar will infer that the echo has come from its most recent pulse, putting the scatterer much closer than in reality. This is referred to as second-trip echoes and usually is detectable because the storm that is beyond  $r_a$  appears elongated and unrealistic. The PRT also governs the maximum unambiguous velocity, given by

$$v_a = \pm \frac{\lambda}{4T_s} \quad (2.2)$$

where  $\lambda$  is the wavelength of the radar. If a target is moving faster than  $v_a$ , the phase shift will be such that the PRT cannot adequately sample the phase and therefore the velocity will be aliased. For the same PRT, higher wavelength radars, e.g. 10-cm S-band radars such as the WSR-88D, have a higher  $v_a$  than shorter wavelength radars such as X-band or C-band radars. Combining Equations 2.1 and 2.2 by solving for  $T_s$ , we end up with the equation

$$v_a r_a = \frac{c\lambda}{2} \quad (2.3)$$

where the RHS of the equation is a constant for a given radar. This means that if one wants to increase the unambiguous velocity, the unambiguous range will decrease and vice versa. This is known as the Doppler Dilemma. An appropriate PRT must be chosen such that there is a good balance between  $r_a$  and  $v_a$  depending on the situation.

Once the radar sends out a pulse, it switches into listening mode for returned echos. The power returned by a point target is given by the radar equation for point targets,

$$P_r = \frac{P_t g^2 \lambda^2 \sigma_b f^4(\theta, \phi)}{(4\pi)^3 r^4 l^2} \quad (2.4)$$

where  $P_t$  is the power transmitted,  $g$  is the gain of the antenna,  $\sigma_b$  is the backscatter radar cross section,  $r$  is the range to the target, and  $l$  is the losses due to attenuation.  $f^4(\theta, \phi)$  has a value ranging from 0 to 1 and has to do with the illumination pattern of the beam and where the target is in relation to the center of the beam. If the target is exactly in the center of the beam, then the value is 1. Notice that  $r$  in

the denominator is to the  $4^{th}$  power, and therefore the returned power decreases drastically as the target gets farther away from the radar. However, for weather radar applications, it is never the case where a single point target must be measured by the radar. Rather, it is much more useful to measure a volume of scatters such as a volume of raindrops together. This introduces the radar equation, which is similar to Equation 2.4 but has very important differences. Mathematically, it is written as

$$P_r = \frac{P_t g^2 \eta c \tau \pi \theta^2 \lambda^2}{(4\pi)^3 r^2 l^2 16 \ln(2)} \quad (2.5)$$

where  $\theta$  is the beamwidth in radians and  $\tau$  is the pulse length. For a spherical water droplet,  $\eta$  is given by  $\eta = \frac{\pi^5}{\lambda^4} |K_w|^2 \int_0^\infty D^6 N(D) dD$ , where  $|K_w|^2$  is the dielectric constant for water  $\approx 0.92$ ,  $D$  is the diameter of the drop and  $N(D)$  is the number of drops in the volume. One of the most important distinctions between Eqn. 2.4 and Eqn. 2.5 is that for the radar equation, the  $r$  on the bottom is no longer to the  $4^{th}$  power but is only squared. Therefore, the loss in power with range is not as significant when expressing the radar equation in terms of a volumed sum.

Both radar equations take into account the backscatter cross section of the targets  $\sigma_b$ , which depends on both the orientation of the target in relation to the radar as well as its size relative to the wavelength of the radar. For example, a piece of paper that faces flat side to the radar would obviously be detectable by the beam, but if the paper were thin side facing the radar, the backscatter cross section would be nearly indetectable. In terms of the size, there are three main regimes that

change the equation of  $\phi_b$ . For a spherical raindrop with diameter  $D$ , if  $D > \lambda$ , then the raindrop is considered to be in the optical regime and

$$\phi_b \approx \frac{\pi D^2}{4} \quad (2.6)$$

which means  $\phi_b$  is dependent on the size of the drop. If  $D \leq \lambda/16$ , then the drop is in the Rayleigh regime and

$$\phi_b = \left(\frac{\pi^5}{\lambda^4}\right) |K_w|^2 D^6 \quad (2.7)$$

where  $|K_w|^2 \approx 0.92$  for water. This is the most common regime for raindrops. Notice that  $\phi_b$  is again dependent on the size of the scatterer. However, for  $D \sim \lambda$ , the drops are in the Mie regime and are much more complicated. The EM wave actually goes to the back of the raindrop and wraps back around to the front and may interact constructively or destructively to the return wave, and therefore it is extremely difficult to relate the power returned to the size of the scatterer since they are no longer related in a simple way. The scattering regimes are summarized graphically in Figure ??.

## 2.1 Dual-Polarimetric Variables

Similar to single-polarization radars, dual-polarization radars measure three moments - radar reflectivity factor ( $Z_H$ ), Doppler velocity ( $v_r$ ), and Doppler spectrum width (W) - but in both the horizontal and vertical polarizations. This additional backscattering in the vertical polarization provides measurements of  $Z_H$ ,  $Z_V$ ,  $V_H$ ,

$V_V$ ,  $W_H$ , and  $W_V$ . With the use of amplitudes and phases of returned echoes, additional variables can be calculated providing information about the characteristics of the scatterer such as type, shape, size, concentration, and orientation. These variables include differential reflectivity ( $Z_{DR}$ ), differential propagation phase shift ( $\phi_{DP}$ ), specific differential phase ( $K_{DP}$ ), and correlation coefficient ( $\rho_{hv}$ ). Each of these variables will be discussed in context of physical interpretations and then applied to real-world examples in section III.

### 2.1.1 Radar Reflectivity Factor

The most fundamental use of radars is the ability to detect areas of precipitation. As discussed previously, this is done by transmitting EM pulses which then interact with surrounding scatterers, such as raindrops. For dual-polarization radars, both the horizontal,  $Z_H$ , and vertical,  $Z_V$ , radar reflectivity factors are measured. For a particular volume,  $Z$  is dependent on both the concentration of scatterers as well as the size distribution. Mathematically, it is expressed as

$$Z = \int_0^{\infty} N(D) D^6 dD$$

where  $N(D)$  is the number of scatterers within the sampling volume and  $D$  is the diameter of the scatterer. The pure units of  $Z$  are  $mm^6m^{-3}$ . However, operationally,  $Z$  is most often measured on a logarithmic scale in units of decibels of  $Z$ , or  $dBZ$ . Because  $Z$  is a variable that relies on the returned power, it is affected by attenuation and therefore must be used with caution when analyzing areas of heavy precipitation.

Users must be especially careful of this when using radars with shorter wavelengths, such as X-band radars, as attenuation problems are magnified for shorter wavelength radars.

### 2.1.2 Differential Reflectivity Factor

Differential reflectivity, defined as

$$Z_{DR} = 10 \log_{10} \left( \frac{Z_H}{Z_V} \right),$$

is a measure of the reflectivity-weighted axis ratio of the target given in units of dB (Seliga and Bringi 1976). In terms of cross section, the diameter of a scatterer aligned in the horizontal is defined as  $a$  and similarly, the diameter of a scatterer aligned in the vertical is defined as  $b$ . From Equation 2, targets with high  $Z_{DR}$  indicate that  $Z_H > Z_V$ , i.e.  $a > b$ , and thus high  $Z_{DR}$  represents scatterers that are aligned in the horizontal, such as oblate raindrops. Therefore, targets with their major axis aligned in the horizontal (vertical) polarization produce positive (negative)  $Z_{DR}$  while targets that return equal power in both the horizontal and vertical polarizations produce a  $Z_{DR}$  of 0 dB.

### 2.1.3 Differential Propagation Phase Shift and Specific Differential Phase

The EM waves that are emitted in the horizontal and vertical polarizations are inherently affected by scatterers through propagation and phase shifts. If the pulse

travels through clear air, then the horizontal and vertical waves travel at the same speed and there is no difference in their phase. However, once the EM pulse passes through a scatterer, such as a raindrop, then the pulse is affected; more specifically, the pulse slows down. If the a scatterer is spherical, then this "slow down" is the same in both the horizontal and vertical directions and there is again no difference in the horizontal and vertical phases after passing through the raindrop. However, if the raindrop is oblate, then the horizontal pulse will be slowed down more than the vertical pulse, and therefore the horizontal phase "lags" behind the vertical phase by a certain degree. This is known as the differential propagation phase shift, or  $\phi_{DP}$ . If the horizontal pulse lags behind the vertical pulse, then  $\phi_{DP} > 0$ , and vice versa. As the scatterer size in a volume increases,  $\phi_{DP}$  also tends to increase.  $\phi_{DP}$  is also proportional to the concentration of scatterers within the volume.

One advantage to using  $\phi_{DP}$  is that since it is a phase measurement, it is not affected by attenuation, partial beam blockage, calibration of the radar, among others unlike  $Z_H$  and  $Z_{DR}$ . Therefore, one of the most important applications for  $\phi_{DP}$  is its usefulness in attenuation correction as well as its capability of providing quantitative precipitation estimation (QPE). However, one drawback to using raw  $\phi_{DP}$  data is that  $\phi_{DP}$  is not a very intuitive variable. In order to get information from a particular part of the storm using  $\phi_{DP}$ , one must actually decipher the way in which  $\phi_{DP}$  is *changing* in order to get a sense of which phase, horizontal or vertical, is slowing down faster. Thus, most of the time meteorologists prefer to use  $K_{DP}$ , which is the derivative of  $\phi_{DP}$  and has units of  $^{\circ}/\text{km}$ . This gives a better sense of how the phase shift is changing in the radial direction. Therefore, regions

of heavy precipitation are collocated with regions of high  $K_{DP}$ . However,  $Z_H$  and  $K_{DP}$  behave much differently in certain situations, such as in the presence ice and frozen particles. This is discussed in more detail below.

#### **2.1.4 Correlation Coefficient**

The correlation coefficient  $\rho_{hv}$  is an indication of the diversity of scatters in a sampling volume as measured by the co-polar correlation between the returned signals in the horizontal and vertical polarizations. In other words,  $\rho_{hv}$  measures the similarity of return horizontal and vertical pulses between the different scatterers within the sampling volume. These signals provide information about the physical characteristics of the volume of scatterers, including type, shape, and orientation. Thus, for a volume of scatterers consisting of a variety of scatterers (e.g., a mixture of hail and rain)  $\rho_{hv}$  would be low. On the other hand, more uniform scatterers within a sampling volume would lead to a high  $\rho_{hv}$  value (i.e. closer to 1). An advantage of  $\rho_{hv}$  is that it is not dependent on particle concentration and/or size and is also resistant to attenuation, beam blockage, and miscalibration of the radar.

## **2.2 Supercells**

### **2.2.1 Environmental Characteristics**

Environmental conditions have a profound influence on convective storm modes. Factors such as instability, wind shear, moisture and forcing/lift are among the

leading contributors to determine storm mode(s). Slight modifications of the environment (small- or large-scale) can drastically alter forecasts and be the difference between a cluster of thunderstorms or discrete supercells. Each of the different storm modes are capable of producing various hazards but vary in degrees of severity. One of the most notorious of the storm modes are supercells because of its overall strength and intensity, as well as life-threatening impacts. Typical hazards for supercells include damaging winds, large hail, torrential rain, and occasionally tornadoes. The organization, maintenance and severity of a supercell are dependent on the deep, persistent mesocyclone. There are three classifications of supercells: classic, high-precipitation (HP), and low-precipitation (LP), each of which are dependent on environmental conditions, as well as two terminologies describing the evolution of supercells: cyclic and non-cyclic ().

The environment favorable for supercell development relies on several factors. Supercells in particular are very susceptible to variations in environmental wind shear, both in magnitude and direction. Firstly, the deep layer wind shear vectors should be perpendicular to the boundary otherwise, conditions are not favorable for maintaining discrete convection and instead favor a multicellular mode. Secondly, the vertical wind field should veer with height which promotes dominant right-moving supercells (discussed more in the next section) and its magnitude should be on the order of 35-40 kts or greater. Additionally, for tornadic supercells, strong low-level (0-1 km) wind shear is beneficial, typically on the order of 15-20 kts or greater. Supercells also require decent instability for explosive growth usually denoted by MLCAPE greater than  $1000 \text{ J kg}^{-1}$ . However, tornadoes have occurred in lower



Figure 2.1: Adapted from Markowski (2002).

CAPE environments as a result of very strong vertical wind shear. High CAPE environments are supported by steep mid-level lapse rates, ample moisture and relatively low dewpoint depressions. In order for supercells to initiate, there must be strong forcing usually provided by a boundary such as a cold front or dryline. There also must be relatively low CIN which can be eroded by an increase in moisture and temperature and broad synoptic-scale lift.

### **2.2.2 Structure and Dynamics**

Figure 2.1 shows a diagram of an idealized supercell with the locations of the rear flank downdraft (RFD), forward flank downdraft (FFD), updraft, and gust front

with associated streamlines. Warm, moist air is drawn into the updraft and is denoted by a notch in the gust front. The FFD is typically located to the northeast of the updraft and usually contains the heaviest precipitation region. As the mesocyclone matures, some of the precipitation from the FFD is cyclonically wrapped around the updraft and descends to the southwest of the updraft, resulting in the RFD. The cyclonic wrapping of precipitation is what gives the supercell its classic kidney bean shape when viewed on radar. Not only can the RFD contain extremely strong winds, but its characteristics have been found to be important for tornado-genesis mechanisms (Lee et al. 2012; Skinner et al. 2014; Mashiko 2016). Note that in HP supercells the precipitation in the RFD may wrap completely around the updraft while in LP supercells the RFD may not have any precipitation. In an ideal case, the gust front would remain relatively close to the precipitation cores; if the gust front surges too far ahead of the precipitation, the warm, moist air feeding the updraft may be cut off.

The FFD is the result of the descent of rain-cooled air. This descent is due to both evaporation and melting of hail as well as precipitation loading. As the forward flank gust front propagates, warm, moist air is lifted which helps to maintain precipitation in the FFD. Additionally, the temperature contrasts between the warm inflow and the cool air within the FFD baroclinically-induced low-level horizontal vorticity along the boundary. In fact, magnitude of baroclinic vorticity can match or be greater than environmental vorticity. The updraft then tilts the low-level baroclinic horizontal vorticity which produces significant low-level vertical vorticity and enhances mid-level vertical vorticity. Thus, low-level mesocyclones usually await

formation of the FFD baroclinic zone. However, this is still not a sufficient condition for the formation of tornadoes since tilting cannot produce vertical vorticity at the surface. Thus, tornadoes may rely on a downdraft to bring vorticity to the ground, a possible role of the RFD.

Though there have been many studies done on understanding the role of the RFD, the complex role of this region has yet to be found. The presence of the RFD is due to the combination of thermodynamic and dynamic effects. Thermodynamically, evaporative cooling from cyclonically wrapping precipitation leads to negative buoyancy and descending air. Dynamically, the updraft acts as a solid column which blocks upper-level flow leading to a downward-oriented vertical pressure gradient force that forces air to descend. If the dynamic processes outweigh the effects of thermodynamic processes in the maintenance of the RFD, the RFD winds at the surface may actually be warmer than its surroundings due to warming during adiabatic descent. The thermodynamic properties of the RFD winds near the surface can have a significant impact on tornadogenesis (Markowski et al. 2002; Grzych et al. 2007). If the RFD winds are able to wrap back around the updraft near the surface and the air is sufficiently buoyant, the stretching of near surface vorticity may occur, a precursor to tornadogenesis.

The updraft is a region of little to no precipitation as a result of high vertical velocities, excluding some HP supercell cases. On radar, the updraft is discernible at upper levels by a bounded weak echo region (BWER) due to the lack of precipitation. The rotating updraft in supercells typically maintains vertical vorticity on the order of  $10^2 s^{-1}$  for a sufficient period of time. The vertical vorticity in the midlevels

is generated by the tilting of streamwise, horizontal vorticity created by the ambient wind shear. Once the updraft intensifies, stretching also plays an important role in the maintenance of vertical vorticity. At early stages in the updraft lifecycle, there are two terms that govern the behavior of the updraft: the linear and nonlinear forcing terms. When the ambient wind shear lacks directional variation, the nonlinear dynamic forcing term dominates which leads to stacked high and low pressures (Figure 2.2). Thus, in this case, the updraft does not favor one propagation direction over another, leading to a supercell split of a right and left-mover. When there is significant directional shear, the linear dynamic forcing term dominates and the high and low pressure centers align in such a manner that rightward propagation of the updraft is favored.

Evolutionarily, supercells often have more than one updraft, a process known as cyclic mesocyclogenesis. This process is the result of the interactions between the updraft and the surging RFD. When the RFD and associated gust front surge outwards, the original midlevel mesocyclone occludes and strengthens due to tilting of the baroclinically generated horizontal vorticity and stretching of streamwise vertical vorticity from descending RFD air. Simultaneously, the surging gust front forces development of a new updraft downshear, resulting in a two-celled updraft structure. As the gust front progresses, the downshear updraft continues to propagate further downshear while the warm air inflow into the original updraft is undercut. Thus, the new updraft replaces the original mesocyclone. The process of cyclic mesocyclogenesis can repeat as long as the RFD continues to surge and occlude the mesocyclone.

## 2.3 Tornadoes

### 2.3.1 Characteristics and Impacts

A tornado is defined as a rapidly rotating column of air that extends from the base of a cloud to the ground. This rotating column is often visible due to condensed water vapor and/or dust and debris particles that surround the column, i.e. the condensation funnel. There are a variety of shapes and sizes that describe the silhouette of a tornado, most commonly used are wedge, stovepipe, elephant trunk, and rope. Most tornadoes are less than 2 km in width, although there are exceptions to this, and vary on a timescale between seconds to hours. Though the majority of tornadoes are produced by supercells, less than 30% of supercells produce tornadoes (Trapp et al. 2005).

The Fujita scale was developed in 1971 by Tetsuya Fujita to systematically rate tornadoes based on estimated wind speeds and associated damage, but it was not until 1973 that the scale became operational nationwide. However, research years later found that the wind speeds used in the Fujita scale were grossly overestimated and inaccurate (WSEC 2006). It was revised in 2007 and is now referred to as the Enhanced Fujita (EF) scale (see Table 2.1). This scale more accurately associates the wind speed estimations to damage, and also takes into account structural integrity.

The rating is accredited by the National Weather Service after conducting a thorough damage survey that identifies Damage Indicators and Degrees of Damage along the damage path. The damage is assessed and assigned a range of wind speeds

Table 2.1: Fujita scale and Enhanced Fujita scale with associated 3 second wind gusts.

| <b>F-Rating</b> | <b>3-s Wind Gust (mph)</b> | <b>EF-Rating</b> | <b>3-s Wind Gust (mph)</b> |
|-----------------|----------------------------|------------------|----------------------------|
| F-0             | 40-72                      | EF-0             | 65-85                      |
| F-1             | 73-112                     | EF-1             | 86-110                     |
| F-2             | 113-157                    | EF-2             | 111-135                    |
| F-3             | 158-206                    | EF-3             | 136-165                    |
| F-4             | 207-260                    | EF-4             | 166-200                    |
| F-5             | 261-318                    | EF-5             | $\geq 201$                 |

that best estimates the tornado strength at a given location. The upper limit of this range along the damage path is used to determine the overall rating of the tornado.

From 1985 to 2017, the United States averaged 1,161 tornadoes and 68 fatalities per year (SPC 2016). The number of fatalities increase with EF scale up to an EF-3 and then decrease for EF-4 and EF-5 . However, this decrease at the upper end of the scale can be attributed to the rarity of violent (EF-4+) tornadoes. In fact, although violent tornadoes account for only 0.6% of all tornadoes, they are responsible for just over 50% of all tornado-related fatalities. The rarity of EF-5 tornadoes is so extreme that there have only been 20 occurrences in 12 different years since 1985. On the lower end of the scale, weak (EF-0 and EF-1) tornadoes account for 88.4% of all tornadoes but are responsible for only 5.7% of fatalities. Economically, tornadoes have one of the highest monetary impacts among all weather-related disasters. In any given year, tornadoes account for 36% of all insured losses since 1983 (Hartwig 2014). From 1993 to 2012, the total monetary impact of tornadoes was \$140 billion and in 2011 alone, amounted to \$26 billion in tornado damage.

There are two types of tornadoes: Type I and Type II. Type I tornadoes are associated with a parent supercell and its accompanying mesocyclone. Type II tornadoes are not associated with a mesocyclone, and are generally much weaker and short-lived. The latter are commonly known as landspouts, gustnadoes or waterspouts. Studies have shown that Type II tornadoes are formed predominantly by the stretching of pre-existing vertical vorticity generated by localized wind-shift lines (Bluestein and Jain 1985; Wakimoto and Wilson 1988; Brady and Szoke 1988). Enhanced areas of vertical vorticity create localized areas of low pressure which can induce an updraft to form. If the strength of the updraft is sufficient (and is collocated with the vertical vorticity), the pre-existing vertical vorticity will be stretched to form a tornado. This is why landspouts and waterspouts are commonly observed beneath cumulus clouds. Unlike Type II tornadoes, the mechanisms for Type I tornadoes are not as well-understood and there is still much debate on the origins of near-surface vertical vorticity. This will be discussed more in-depth in the next section.

### **2.3.2 Tornadogenesis Mechanisms**

There are three distinct stages in tornadogenesis. The first stage is the formation of the mid-level mesocyclone which is well-understood and as stated before comes from the tilting of horizontal vorticity generated by environmental wind shear. Stage two of tornadogenesis is not well-understood and consists of the generation of near-surface vertical vorticity. It is hypothesized that the RFD plays an important role in

either the tilting of horizontal vorticity into the vertical or by simply advecting pre-existing vertical vorticity downward. Stage three of tornadogenesis is the formation of the tornado and is generally accepted as the stretching of the near-surface vertical vorticity by an intense updraft (Ward 1972; Walko 1993).

There are a handful of theories to the generation of near-ground rotation. One of the most popular theories of bringing strong vertical vorticity aloft to the surface is the dynamic pipe effect (DPE) (Leslie 1971). In the DPE, there is flow that is in cyclostrophic balance associated with the low-level mesocyclone. However, just below the flow in cyclostrophic balance, there is convergence and associated rising motion that acts to increase the vertical vorticity at this level by stretching. After the convergence acts on the vertical vorticity for a certain amount of time, this level reaches cyclostrophic balance as well. This process continues until the vortex reaches the surface. Empirical evidence of the DPE was presented in the 24 May 1973 Union City, Oklahoma tornado. In this case, a tornado vortex signature was identified at mid-levels first and built downward with time (Brown et al. 1978). However, other observations and simulations have shown that the vortex either forms simultaneously through all levels or from the bottom-up (Vasiloff 1993; Trapp and Fiedler 1995).

Because of this discrepancy, two modes of tornadogenesis have been proposed. Mode I is when the tornado forms aloft and is brought down to the surface via a mechanism such as the DPE. In Mode II of tornadogenesis, the tornado forms at the surface extremely rapidly. From an operational perspective, Mode I tornadoes are easier to warn since the development of a TVS aloft proceeds the formation of a tornado. Unfortunately, recent studies using WSR-88Ds and rapid-scanning

mobile radars have failed to find any evidence of Mode I tornadogenesis occurring (Alexander 2010). The lack of findings of Mode I tornadogenesis may be due to insufficient temporal resolution especially in regards to the WSR-88Ds and lack of observations of mid-level mesocyclone using mobile radars. However, it is also possible that most tornadoes form via Mode II tornadogenesis.



Figure 2.2: Adapted from Markowski and Richardson (2010).

## Chapter 3

### Data and Methodology

Radar variables were converted to a Cartesian grid from polar variables.

# Chapter 4

## Overview

plots (zh, zdr, kdp, rhohv, vr) possible DRCs?

hc / *hc<sub>hail</sub>plotsareatimeseries*

enhanced zdr/kdp centroids plots area time series

maximum delta v KOUN

vertical plots 3d plots

zdr column / arc per Van Den Broeke

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