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POLARIMETRIC DOPPLER SPECTRA IN SOUTHERN PLAINS TORNADOES  
COMPUTED FROM MOBILE RADAR DATA

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COMPUTED FROM MOBILE RADAR DATA

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BY THE COMMITTEE CONSISTING OF

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# Abstract

Doppler wind velocity estimates depend on the assumption that scatterers in the atmosphere are moving at the same velocity as their surrounding air molecules, as the radio waves used in weather radars do not significantly backscatter off air molecules. Most of the time, this assumption is valid, but in tornadoes, wind speeds become high and the airflow exhibits substantial curvature. Larger and more massive scatterers accelerate more slowly under the force of the wind and are also subject to an increased centrifugal force. Thus, these scatterers do not move at the same velocity as the wind, and velocity estimates can be subject to large errors/biases. In order to study tornado dynamics, errors in wind fields must be diagnosed and minimized. This can be accomplished by spectral analysis of raw, dual-polarization, in-phase and quadrature (I/Q) data. Since 2018, efforts have been made using the University of Oklahoma’s rapid-scan, X-band, polarimetric, mobile Doppler radar to collect I/Q data in tornadoes and verify these velocity errors. Through Fourier analysis of I/Q signals, dual-polarization spectral densities (DPSDs) can be computed, which leverage the distinct polarimetric characteristics of rain and debris to separate their respective motions.

In 2023 and 2024, RaXPol collected I/Q data from within three tornadic vortices, with the last one, on May 23, 2024, being the most comprehensive. This dataset was collected in an EF-2 tornado using a scanning strategy involving multiple low-level scans through the debris cloud of the tornado. The radar employed a slow azimuthal rotation ( $6^\circ \text{ s}^{-1}$ ) with sector scanning, achieving over 650 pulses per degree of azimuth. The spectra were evaluated for quality using a convergence metric that optimizes parameters for DPSD computation and assesses the requirement for slow scanning. With computed DPSDs and an automated fuzzy logic debris classification algorithm (DCA), spectra were analyzed, and areas of strong velocity bias were identified, consistent with numerical simulations. These results demonstrate the potential for improved tornado dynamics retrieval through spectral polarimetric analysis.

# Chapter 1

## Introduction & Background

Tornadoes contain among the highest observed wind speeds on the planet, with wind speeds estimated by mobile radar to be in excess of 300 miles per hour in the strongest ones. These winds cause extreme damage in narrow but intense corridors. In 2023, tornadoes were responsible for  $\sim$ \$1.38 billion in damages, 86 fatalities, and 918 injuries, despite being a spatially confined phenomenon (NCDC 2024). As a high-impact phenomenon, it is imperative to study the processes that generate, maintain, and dissipate tornadoes in an effort to improve predictability.

All types of convective (single-cell, multi-cell, mesoscale convective system (MCS), and supercell) storms have been documented to produce tornadoes; however, the most intense and long-duration tornadoes have been observed with supercells. Supercells are differentiated from other types of convection by two defining characteristics: a vertically deep, persistent rotating updraft (known as a mesocyclone) and a general long-lived nature (Bluestein 2013).

### 1.1 Tornado Dynamics

The fundamental mechanism for tornadogenesis is the presence of low-level vertical vorticity (either storm-generated or present in the environment) and a mechanism to stretch that vorticity, amplifying wind speed. For cases where environmental vertical vorticity is not present in sufficient quantities to be stretched into a tornado vortex,



a storm is capable of generating its own. Baroclinically generated horizontal vorticity or environmental horizontal vorticity, both oriented in a streamwise direction (storm-relative vorticity vectors pointed towards the updraft of a storm), can be tilted into cyclonic near-surface vertical vorticity by the horizontal gradients in vertical wind in proximity to an updraft (Bluestein 2013).

A balance must be struck between thermodynamic and kinematic quantities within different locations of a given storm for efficient tilting of streamwise vorticity. Downdrafts must not be too cold and prevalent such that gust fronts propagate away from updrafts, resulting in outflow dominance. Conversely, downdrafts that are too weak compared to storm-relative inflow will fail to collect near-surface vorticity to be stretched by the updraft (Nixon et al. 2024; Peters et al. 2023; Fischer and Dahl 2023). For supercells specifically, the rear-flank gust front provides a convergent environment capable of collecting vorticity.

Once a tornado has developed, most conceptual models treat it as axisymmetric, which means the flow structure is symmetric about a central vertical axis. The simplest two-dimensional vortex model orthogonal to the axis of rotation is known as the Rankine vortex. This model features a radius of maximum wind (RMW), inside which the flow field behaves as a solid body vortex and outside which the flow field behaves as an irrotational vortex. The pressure field exhibits a strong low pressure in the center of the tornado, which has been observed with in-situ observations (Samaras and Lee 2006). The flow field is said to be in cyclostrophic balance, where in cylindrical coordinates, the inward-directed pressure gradient force is balanced by the outward-directed centrifugal force. This balance neglects the Coriolis force, due to the small scale.

However, a real tornado contains vertical winds, which the Rankine vortex does not capture because it assumes purely vortical. Thus, while the Rankine vortex is a reasonable conceptual model at higher elevations, this assumption breaks down near

the surface. Here, the flow field can be broken down into 5 main flow regions. Turbulence and surface drag-aware computer model simulations diagnose additional sources of near-surface horizontal vorticity, which may be important for tornadogenesis and maintenance (Rotunno and Bluestein 2024). A large-eddy simulation (LES) that accounts for surface roughness from Lewellen (1993) highlights a shallow surface layer 10 to 100 meters deep, where flow is turned strongly convergent. This boundary layer includes the corner flow region, where this inflow is abruptly lifted.

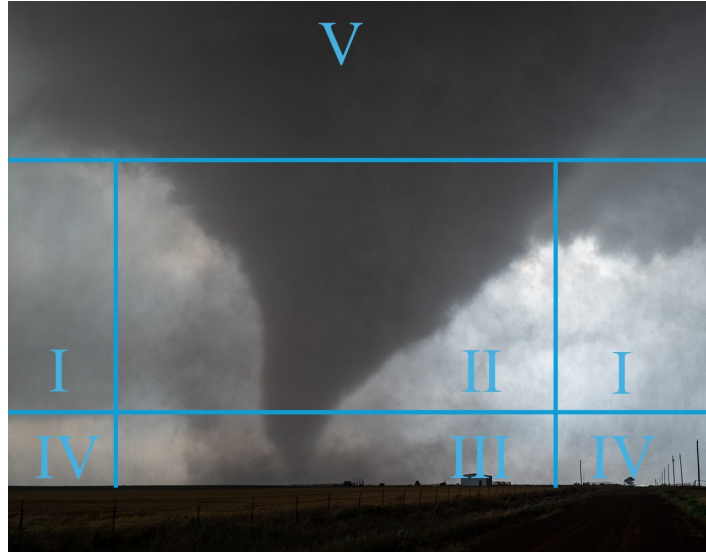


Figure 1.1: Flow regions and recreated figure from (Markowski and Richardson 2010). I) outer core, II) inner core, III) corner flow region, IV) surface layer, and V) updraft. Photographed by Jim Tang.

For the corner flow region, structure can vary with Reynolds number  $Re = UL/\nu$ , where  $U$  is a velocity scale,  $L$  is a length scale and  $\nu$  is viscosity (Reynolds 1883), as well as swirl ratio  $S = V_t/w$ , where  $V_t$  is azimuthal velocity and  $w$  is the mean vertical velocity (Church et al. 1979). For a fixed  $Re$ , as  $S$  increases, vortex structure becomes more complex. At low swirl ratios below 1, where motion is updraft-dominated, a single-cell vortex with an updraft at the core can be observed. At higher swirl ratios, downdrafts may develop in the core of the tornado and even reach the ground, as seen

in Figure 1.2. As the swirl ratio increases further, the symmetric vortex structure will transition into a multiple vortex structure (Davies-Jones 1986). Notably with these downdrafts, areas of wind motion radially outward from the tornado, or divergence, can develop.

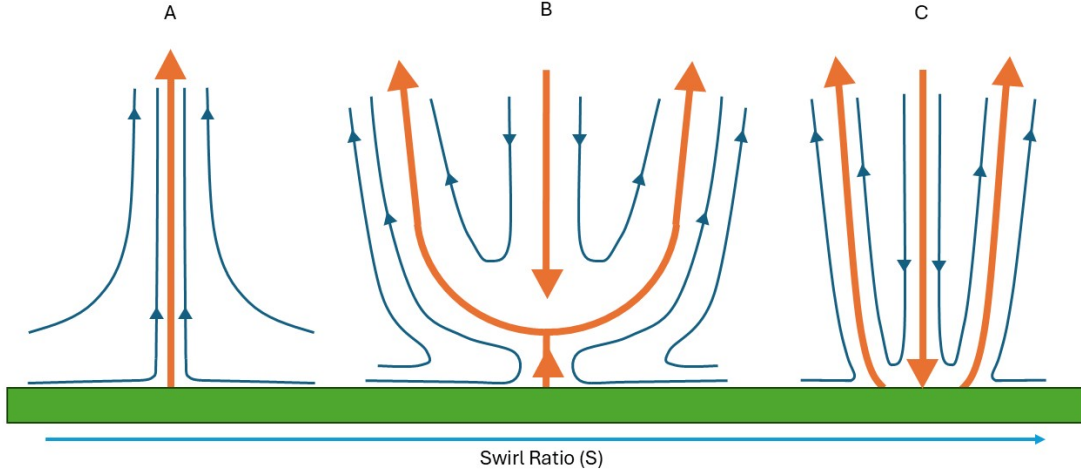


Figure 1.2: Tornado vortex structure at a constant Reynolds number with increasing swirl ratio. Orange curves show updraft/downdraft maxima, while blue curves show irrotational wind streamlines. A) A single-cell vortex with low swirl ratio. B) A Two-cell vortex structure aloft and single-cell vortex structure near the surface as a downdraft begins descending. C) A two-cell vortex structure. As swirl ratio increases further past this point, a multiple vortex structure is observed. Re-created from Rotunno and Bluestein (2024) Figure 1.

More recently, a semi-slip numerical simulation by Fischer and Dahl (2022) outlines a conceptual model where axisymmetric convergent and vortical flow in this shallow surface layer advects streamwise surface-layer drag-induced vorticity toward an ongoing tornado. This vorticity is abruptly tilted into the vertical directly by the corner flow in the tornado, supplying surface layer vertical vorticity to be stretched.

These advancements in the knowledge of tornado dynamics are mainly derived from high-resolution numerical simulations. In non-idealized conditions, tornadoes are likely

not axisymmetric owing to their presence near storm-scale boundaries. Regardless, numerical simulations and observed behavior highlight the need for low-level observations to diagnose kinematic and thermodynamic features of storms and associated tornadoes.

## 1.2 Radar Observations

Given the extreme conditions in tornadoes, direct observations are difficult, if not impossible. Some observations have been made by way of probes deployed in the path of tornadoes (e.g., Karstens et al. 2010). However, our most reliable source of kinematic observations comes from pulsed Doppler radar. Across the United States, a network of 159 stationary radars managed by the National Oceanic and Atmospheric Administration’s (NOAA’s) National Weather Service (NWS) provides continuous observations of the atmosphere, with the primary objective of monitoring meteorological scatterers. This network is known as NEXRAD (Next Generation Weather Radar) (ROC 2025).

Pulsed Doppler radar is capable of determining properties of scatterers (like tornado-lofted debris or rain) in the atmosphere. A pulse of electromagnetic radiation is transmitted into the atmosphere. A fraction of the radiation is backscattered towards the radar, which is received after some time. Most often, a carrier frequency is chosen that will backscatter significantly from larger hydrometeors like rain, but not significantly and possibly immeasurably from cloud droplets. Information such as backscattered pulse amplitude, frequency shift from the carrier wave, and travel time give us information about how many scatterers there are in a volume, their mean radial (from the radar) velocity, and how far they are. If we apply an assumption that scatterers are moving at the same speed as the wind, we can make inferences about wind speeds remotely, without direct equipment being placed inside tornadoes. With this, the NEXRAD network is used to detect tornadoes.

The main benefit of the NEXRAD system is its unparalleled real-time spatial and temporal observational coverage. However, many of these observations are high in the atmosphere, sometimes exceeding 3000 m AGL. As we get farther from the radar, beam height increases due to the curvature of the earth. Additionally, azimuthal resolution decreases due to the circular geometry of radar observations. As determined in Section 1.1, processes important to the formation and maintenance of tornadoes occur extremely close to the ground and are relatively fine-scaled. Beyond detection, fixed-site radars exhibit significant weaknesses in the diagnosis of low-level kinematic properties of tornadoes.

The solution to this problem is the ground-based mobile radar, which, due to flexibility in deployment position, improves the issue of beam height, as well as enables far more fine-scaled observation of storm-scale structures than fixed-site radars (Wakimoto and Martner 1992; Wurman et al. 1996).

Bluestein (2022) gives a brief overview of field projects involving mobile radars that had an objective of studying tornadoes. Initially, the need for mobile radars was identified in the late 1980s. The Los Alamos National Lab portable continuous wave frequency modulated radar was the first in a long series of mobile radars to observe tornadoes, used manual scanning, storing data on audio tapes (Bluestein and Unruh 1989). This radar made the first measurements of F-5 wind speeds in a tornado. The U. Mass. W-band radar was the first automatically scanning, pulsed, truck-based radar. W-band radars use a carrier frequency high enough to discern cloud droplets. However, this frequency range attenuates severely through meteorological scatterers, highlighting the need for a lower frequency capable of discerning features deeper within storms. Following this, the Doppler on Wheels (DOW) X-band radars were developed, becoming an extremely productive source of data detailing the fine-scaled structure of tornadoes. In order to scan volumes more quickly, rapid-scanning techniques were

adopted in mobile radars, including electronically scanning (phased array) techniques. These initial radars were the Rapid-Scan DOW and the MWR-05XP. However, unlike the U. Mass. W-band radar and the DOWs, these rapid-scanning radars did not have polarimetric capabilities, which are necessary to observe tornadic debris signatures. More recently new mobile phased array radar projects have been undertaken by the University of Oklahoma's Advanced Radar Research center, like PAIR (polarimetric atmospheric imaging radar, Salazar et al. 2019) and Horus (Palmer et al. 2023),



Figure 1.3: Photograph of RaXPol in front of a thunderstorm. Photographed by Leah Swinney.

The Rapid-scan X-band Polarimetric radar (RaXPol) (Pazmany et al. 2013) is capable of both rapid-scan and polarimetric observations. As pictured in Figure 1.3, this radar's rapid-scan capabilities have allowed for increased temporal resolution per sweep. The pedestal is capable of rotation at  $180^\circ \text{ s}^{-1}$ , enabling a sweep every two seconds and a volume scan every 15-25 seconds.

Rapid-scan capabilities have enabled the collection of observations that corroborate some findings from numerical simulations. In Houser et al. (2022), observations from seven tornadoes with RaXPol showed evidence of non-descending tornadogenesis in five out of seven tornadoes in the lowest 2 km above radar level (ARL). Thus, in the majority of these observed cases, it is evident that azimuthal shear, which can be used as a proxy for vertical vorticity, is either amplified and propagated upward from the surface or uniformly amplified along the full depth of the low-level vortex. This further highlights the need for accurate low-level observations.

### 1.3 Debris Centrifuging

From a quantitative standpoint, there are issues with the assumption that scatterers move at the same speed as the wind, as implied in Section 1.2. Originally, Snow (1984) described the motion of dust in columnar vortices like dust devils. This paper describes the structure of dust sheaths surrounding axisymmetric vortices and how multiple sheaths may be formed by multiple “puffs” of dust from the surface. These sheaths result from a balance between centrifugal forces and gravity, contradicting the assumption that the dust would be moving at the same speed as the wind.

More recently, debris centrifuging has been studied in numerical models to discern errors between the real wind and the velocity of scatterers. Dowell et al. (2005) conducted such simulations in one and two dimensions assuming axisymmetry, modeling differences between the real wind, small scatterers like rain, and large scatterers like debris or hail. Differences in both radial and azimuthal velocity (defined relative to the tornado’s axis) were studied. For the sake of clarity, throughout this work, “radial velocity” refers specifically to the cylindrical velocity component directed toward or away from a tornado’s axis of rotation, while “Doppler velocity” or “velocity” used in

the context of radar observed measurements denotes the spherical coordinate velocity component directed toward or away from the radar.

Regions of debris motion the tornado vortex analogous to those discussed in Section 1.1 are exposed by numerical simulations: the vortex aloft and the corner flow region near the surface. Rain and debris/large hail exhibit radial velocity biases outward relative to the wind aloft. However, debris demonstrates higher biases due to a larger degree of centrifuging. For azimuthal winds aloft, we see that all scatterers have a negative azimuthal velocity bias, but the effect is again higher for debris/large hail, indicating an increased amount of slowing due to a larger mass and slower acceleration.

In the corner region, within the shallow layer of strong inflow, centrifuging-driven biases are less present for all scatterers, as air motions are more irrotational. This is not only shown by numerical simulations previously mentioned, but damage surveys of an EF-5 tornado that impacted Moore, OK in 2013 showed tree fall patterns that implied highly convergent winds. The damage suggests that the radial wind may have been stronger than the azimuthal wind near the surface (Atkins et al. 2014). Thus, Dowell et al. (2005) find that acceleration-driven biases are also present. For debris/large hail, significant differences emerge in the radial and azimuthal components from the wind. Fast-falling scatterers with small radial and large azimuthal velocity from aloft fall into wind with strong radial and weak azimuthal components with respect to the tornado. In this layer, these scatterers are unable to accelerate before landing on the ground, resulting in a radial bias away from the tornado and an azimuthal bias exceeding wind speed.

In three dimensions, LES simulations determined additional properties of debris in a tornado vortex. Unlike the previous simulations, a coupled approach via a drag force feedback parametrization was used, allowing for debris loading to affect wind speeds, adding a layer of complexity. The exchange of momentum between scatterers and air



molecules tends to lower wind speed magnitudes (Lewellen et al. 2008; Bodine 2014; Bodine et al. 2016a). In simulations run by Bodine (2014, pp. 178–202), 1 mm and 0.25 mm sand particles deviated differently from wind speeds. Aloft, in the radial direction, differences were on the order of  $20 \text{ m s}^{-1}$ , with sand motions radially outward compared to wind. In the azimuthal direction, sand was approximately  $10 \text{ m s}^{-1}$  slower than the wind. This remained true for all simulation runs with varying sand concentrations and particle diameters. However, in the inflow layer, when sand was present, it retained azimuthal velocity from aloft and maintained a positive bias on the order of  $5 \text{ m s}^{-1}$ .

Analysis of tornadic wind fields becomes challenging with these scatterer biases. Since most radars are capable of detecting scatterers the size of raindrops or larger, sampling tornadic winds without such scatterers is impossible. Wakimoto et al. (2012) and Nolan (2013) highlight issues with using radar-based techniques to determine kinematic fields. Wakimoto et al. (2012) conducted a ground-based velocity track display (GBVTD), which is a technique to retrieve two-dimensional or three-dimensional wind fields from single-Doppler data, taking advantage of the axisymmetry of a tornado vortex. Such an analysis for a tornado near LaGrange, Wyoming, on June 5, 2009, yielded an intense axial downdraft through the center of the tornado. This resulted from centrifuging-induced bias in divergence estimates, as well as the radar beamwidth being too wide to isolate the surface inflow layer. After correcting for centrifuging, this strong central downdraft had a much-reduced intensity.

Nolan (2013) explored this idea further, conducting GBVTD on simulated tornado data with known wind velocities. The technique was shown to result in velocity errors of  $5\text{--}10 \text{ m s}^{-1}$ , even without centrifuging effects. When accounting for data that, like typical radar data, did not accurately sample the inflow layer, errors grew, decreasing wind speed and making the tornado wider. Finally, when adding radial centrifuging effects of  $8 \text{ m s}^{-1}$ , biases which are reasonable as per Bodine (2014), likely unphysical

extreme downdrafts were observed. These biases, being intrinsic to the nature of radar observations, may persist even when using advanced retrieval techniques such as multi-Doppler.

These Doppler velocity limitations highlight the need to filter debris-induced biases. Methods to classify debris will be discussed in the following section.

## 1.4 Radar Signal Properties & Processing

Details about radar signal properties and signal processing techniques are outlined in Doviak and Zrnic (1993). This includes information about types of scatterers and scattering regimes, choice of carrier frequency and pulse length, how scanning is done, and how to process raw data from the radar into radar variables.

### 1.4.1 Waveform Properties

Fundamentally, the waveform transmitted by a Doppler radar is defined by a few parameters. The radar has a carrier frequency, pulse length, and pulse repetition time. The carrier frequency determines many properties of how the transmitted radar beam propagates and scatters off particles. From a physical standpoint, the carrier frequency is an important consideration in antenna design. Radar carrier frequencies are classified by “bands”, where a range in frequencies or wavelengths is classified by a letter. Bands for common carrier frequencies are given in Table 1.1.

The carrier wavelength has direct implications on the nature of scattering for meteorological scatterers. For a scatterer, we can define a size parameter  $x$  in Equation 1.1:

| Carrier Frequency Bands |                 |                 |                |
|-------------------------|-----------------|-----------------|----------------|
| Band                    | Frequency (GHz) | Wavelength (cm) | Example Radar  |
| S-band                  | 2–4             | 7.5–15          | NEXRAD WSR-88D |
| C-band                  | 4–8             | 3.75–7.5        | PAIR           |
| X-band                  | 8–12            | 2.5–3.75        | RaXPol         |
| W-band                  | 75–110          | 0.27–0.4        | U. Mass W-Band |

Table 1.1: Common carrier frequencies used by weather radars, adapted from American Meteorological Society (2024).

$$x \equiv \frac{2\pi r}{\lambda} \quad (1.1)$$

In Equation 1.1 (Equation 12.1 in Petty 2006, p. 345),  $r$  represents the radius of a spherical particle, and  $\lambda$  represents the wavelength of incident radiation. For size parameters less than roughly 0.2, the scattering regime is known as Rayleigh. Rayleigh scattering is isotropic, and backscattered radiation increases monotonically with particle size. This can be modeled with a quantity known as the backscatter cross-section ( $\sigma_b$ ).

$$\sigma_b \approx \frac{\pi^5}{\lambda^4} |K_m|^2 D^6 \quad (1.2)$$

$$K_m = \frac{m^2 - 1}{m^2 + 2} \quad (1.3)$$

In equation 1.2 (Equation 3.6 in Doviak and Zrnic (1993, p. 35)),  $D$  represents the diameter of the scatterer, and  $K_m$  represents a complex dielectric factor where  $m$  is the complex index of refraction of the scatterer. For liquid water scatterers,  $|K_m|^2 \approx 0.93$ .

For size parameters beyond 0.2, the assumption in Equation 1.2 breaks down as the Mie scattering regime is entered. Scattering in this regime is no longer isotropic, and

the backscatter cross-section no longer varies monotonically with increasing scatterer diameter.

Our goal with radar observations is for stronger returns to correspond to more or larger scatterers, so it is important that particle sizes commonly found in the atmosphere fall into the Rayleigh regime with our chosen carrier wavelength. This typically includes S-band, C-band, and X-band radars. These wavelengths are such that raindrop-sized and larger scatterers fall within the Rayleigh range, while cloud droplets have negligible measurable returns. For X-band, large hailstones or debris may fall within the Mie regime.

Another issue is attenuation. Equation 1.2 shows carrier wavelength in the denominator, so backscatter cross-sections increase with decreasing wavelengths. However, there is a finite amount of energy available to be scattered or absorbed in a radar pulse. Thus, through precipitation, shorter wavelengths attenuate faster. At X-band wavelengths, attenuation can be so severe that observations through a heavy precipitation core are not possible. Even shorter wavelengths, like W-band, can backscatter cloud droplets but suffer from extremely severe attenuation. Bodine (2014) found that at W-band, if debris loading significantly impacts tornadic wind fields, the transmitted radar signal would be fully attenuated.

Considering attenuation and scattering regimes, longer wavelengths, like S-band, seem ideal. However, a longer wavelength requires a larger dish to transmit a narrower beam.

$$\theta_{1,\text{deg}} = \frac{72\lambda}{D} \tag{1.4}$$

Equation 1.4 is adapted from Equation 3.2b of Doviak and Zrnic (1993, p. 24), where  $\theta_{1,\text{deg}}$  is the half-power (3-dB) beamwidth in degrees,  $\lambda$  is the wavelength of the

carrier frequency, and  $D$  is the diameter of the parabolic dish. By this relation, smaller wavelengths require smaller dishes for equivalent beamwidths. The goal is to bring a radar close to the storm while still collecting observations with high angular resolution. It is impractical to bring an 8.53 m (ROC 2025) dish (required for a  $0.925^\circ$  half-power beamwidth) close to a storm, as a larger beamwidth compromises angular resolution.

For this reason, a compromise must be struck. As discussed in Section 1.2, the most productive close-range observations of tornadoes have occurred with X-band radars like the DOWs and RaXPol, as X-band strikes a balance between wavelength, dish size, and beamwidth. However, for the shallow inflow layer in the corner region, Wakimoto et al. (2012) highlights that observations must either be conducted from a few kilometers away or with a radar operating at a smaller wavelength, like W-band.

For this work, we will analyze data gathered from RaXPol, which operates at a 3 cm carrier wavelength. A 2.4 m diameter dish for this radar allows a half-power beamwidth of  $1.0^\circ$ , comparable to NEXRAD.

The other parameters that define a radar's transmitted waveform are pulse width ( $\tau_s$ , in seconds), and pulse repetition time ( $T_s$ , in seconds; abbreviated PRT) or pulse repetition frequency (PRF, in Hz). Pulse length presents a trade-off between transmitted power and range resolution. A longer pulse means greater transmitted power, allowing improved sensitivity to smaller returns. However, this increases ambiguity as to where a given scatterer is within the pulse width, worsening range resolution.

PRT presents a characteristic trade-off known as the Doppler dilemma. The radar must be pulsed in order to determine the location of scatterers. Based on the time delay of the backscattered pulse, using the known speed of light, we can determine the distance to the scatterers. A longer PRT increases the maximum unambiguous range.

$$R_a = \frac{cT_s}{2} \tag{1.5}$$

In Equation 1.5,  $R_a$  represents the maximum range the radar can determine a scatterer's location before it becomes ambiguous whether it was detected by the subsequent pulse, and  $c$  represents the speed of light. If returns are originating from beyond  $R_a$ , they are interpreted as being returned from the next pulse in a phenomenon known as range folding.

$$v_a = \frac{\lambda}{4T_s} \quad (1.6)$$

Due to the Nyquist sampling theorem, which implies that to sample a wave, one must sample twice the highest frequency, a low PRT limits the maximum unambiguous (Nyquist) velocity, as shown in Equation 1.6. Real velocities exceeding  $v_a$  will appear aliased, as seen in Figure 1.4. Note that  $v_a$  also depends on the carrier wavelength, demonstrating an advantage of longer wavelengths. The reason for this limitation will be discussed in the next section. When selecting a waveform, the PRT represents a trade-off between minimizing velocity aliasing and avoiding range folding.

## 1.4.2 Radar Variables

### 1.4.2.1 I/Q time series

The raw data recorded from the hardware signal processing system is known as in-phase/quadrature (I/Q) data. The transmitted signal must have a carrier frequency, which has all the properties discussed in Section 1.4.1. However, to convert to a digital signal, the hardware must be able to sample the analog signal at discrete intervals. Sampling the original carrier frequency is impractical due to hardware constraints, as it requires a high rate of sampling and a large amount of storage due to the Nyquist theorem limitations. To address this, the received signal is mixed with the original

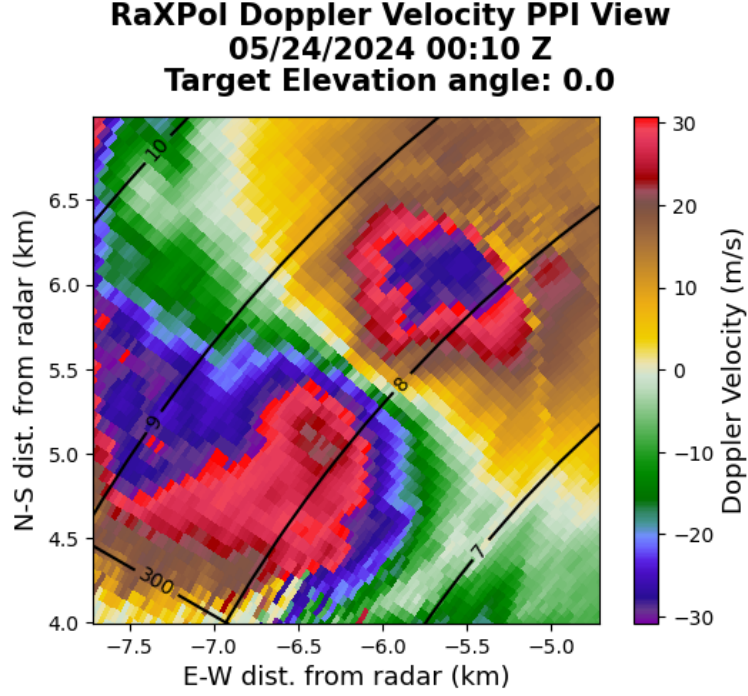


Figure 1.4: RaXPol Doppler velocity PPI on 05/24/2024 at 00:10 UTC. With a Nyquist velocity of  $30.8 \text{ m s}^{-1}$ , velocities higher than this go to the other side of the spectrum.

local oscillator signal and filtered through an analog downconversion chain in order to remove the carrier frequency. This yields a baseband complex signal, which encodes the amplitude of the returned signal in the complex magnitude and the phase shift relative to the original signal in the complex argument.

$$V(t) = I(t) + jQ(t) \quad (1.7)$$

$$I(t) = \sum_{i=0}^N \frac{|A_i|}{\sqrt{2}} \cos(\gamma_i) \quad (1.8)$$

$$Q(t) = - \sum_{i=0}^N \frac{|A_i|}{\sqrt{2}} \sin(\gamma_i) \quad (1.9)$$

$$\gamma_i = \frac{4\pi r_i}{\lambda} + \frac{4\pi v_{r,i}}{\lambda}t - \psi_{(t+s),i} \quad (1.10)$$

The above equations give the form of I/Q data from a continuous wave radar within a given resolution volume where  $j$  is the imaginary unit, and  $t$  is time. In Equations 1.8 and 1.9,  $i$  represents an index of  $N$  scatterers in a resolution volume.  $A_i$  is the amplitude of the backscattered signal from the  $i$ th scatterer, and  $\gamma_i$  is the  $i$ th scatterer phase. In Equation 1.10,  $r_i$  is the range of the  $i$ th scatter from the radar,  $v_{r,i}$  is the radial velocity of the  $i$ th scatterer away from the radar, and  $\psi_{(t+s),i}$  is the phase shift from the transmitter and scattering effects.

For pulsed radar, this signal is multiplied by a pulse function, which approximates a binary function (0 or 1). Pulsing allows the radar to determine the range of scatterers. The continuous received signal is digitized and stored if I/Q data collection is enabled. After removing the carrier frequency, any frequency shifts that were imposed by meteorological scatterers (raindrops moving radially with respect to the radar) become the only remaining frequency present in the baseband signal.

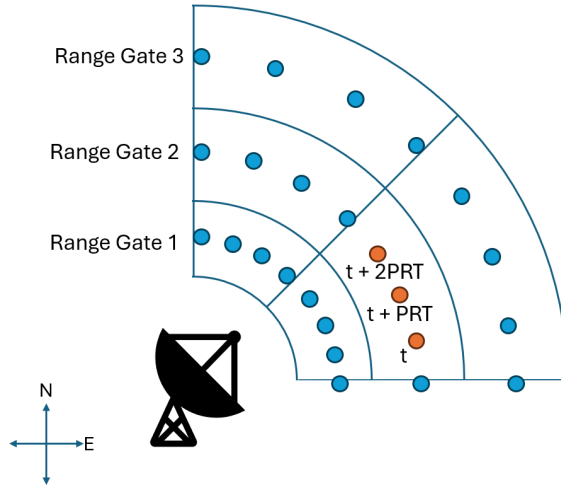


Figure 1.5: Schematic showing pulse grouping strategy to form an I/Q time series. The pulses within a given length in azimuth at a specific range make up a dwell.



We can construct a discrete I/Q time series  $V(m)$  of length  $M$ , where the time at index  $m$  from the start is given by  $t = mT_s$ , as shown in Figure 1.5. An I/Q data point is obtained for each pulse at every range gate, determined by the sampling frequency. I/Q data points at the same range gate from nearby pulses backscatter off approximately the same hydrometeors within a similar resolution volume defined by beamwidth and range resolution. With an assumption that scatterers in this volume are wide-sense stationary, i.e., their statistics remain approximately constant over the timescale of one PRT, we can extrapolate various properties about the scatterers by constructing an I/Q time series. The group of  $M$  pulses used to construct the time series is known as a “dwell”.

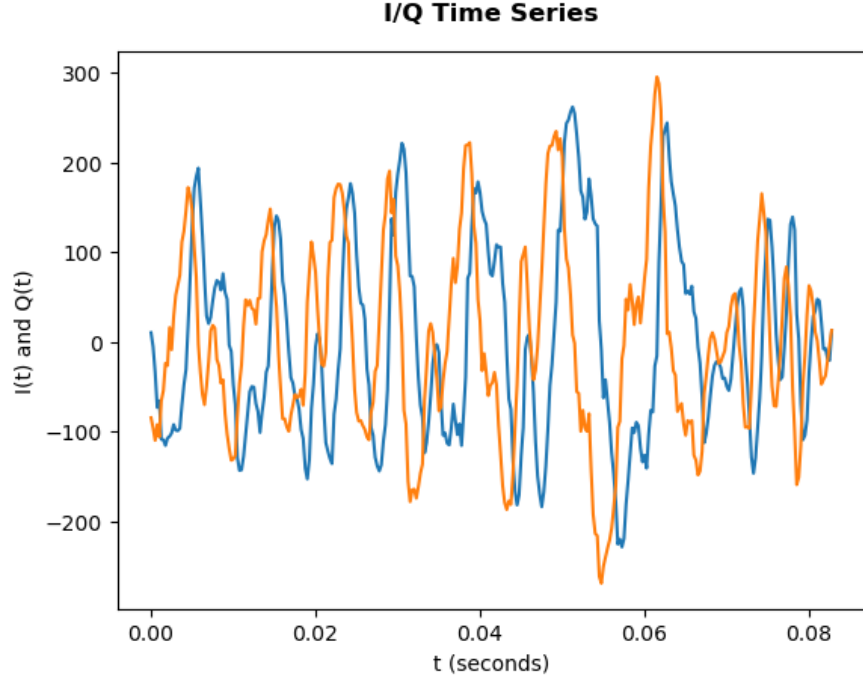


Figure 1.6: Example I/Q time series from RaXPol in the forward flank downdraft of a supercell. The oscillation is the Doppler frequency.

The above time series in Figure 1.6 has a clear frequency and amplitude. This amplitude corresponds to the volume backscatter cross-section, while the frequency corresponds to the velocity of scatterers.

#### 1.4.2.2 Equivalent Reflectivity Factor

The cross-correlation function is defined in Equation 1.11, for any two continuous time series  $X_{1,2}$ , where  $l$  is the time lag and  $E[X]$  is the expected value of a random variable  $X$ .

$$R(X_1, X_2, l) \equiv E[X_1(t)X_2^*(t-l)] \quad (1.11)$$

A discrete cross-correlation function estimator can be defined as the following, for any two discrete time series  $X_{1,2}$  of length  $M$  and index  $m$ , where  $l$  is now the discrete time lag.

$$\hat{R}(X_1, X_2, l) \equiv \frac{1}{M - |l| - 1} \sum_{m=0}^{M-|l|} X_1(m)X_2^*(m-l) \quad (1.12)$$

The autocorrelation function (ACF), noted as  $R_{XX}(X, l)$ , is a specific case of this function where the cross-correlation estimator is applied to a single time series against itself. Given a single dwell I/Q time series  $V$ , the lag-0 autocorrelation results in an estimate of average power returned across the dwell, proportional to amplitude squared.

$$\hat{R}_{XX}(V, 0) = \frac{1}{M} \sum_{m=0}^{M-1} V(m)V^*(m) = \frac{1}{M} \sum_{m=0}^{M-1} I^2(m) + Q^2(m) = \hat{P} \quad (1.13)$$

After subtracting a noise estimate from  $\hat{P}$ , we are left with  $\hat{S}$ , the signal power. We can integrate Equation 1.2 over the full drop size distribution  $N(D)$ . Thus,

$\int_0^\infty D^6 N(D) dD$  is a physical quantity describing our resolution volume and incorporating drop size distribution. Next, we define a quantity called equivalent reflectivity factor  $Z$ , or more simply reflectivity, where  $Z$  is defined as  $\int_0^\infty D^6 N(D) dD$  under the assumption that the resolution volume contains only liquid water scatterers, even when this assumption does not hold. Thus, our estimate of  $Z$  is given in Equation 1.14, ignoring attenuation, where  $\hat{Z}$  is the estimated equivalent reflectivity factor,  $r$  is the range of the I/Q time series  $V$ , and  $C$  is a calibration constant which encodes information about transmitted power, antenna gain, pulse length, beamwidth, and  $|K_w|$ , and  $\hat{S}$  is our estimated signal power.

$$\hat{Z} = \hat{S} r^2 C \quad (1.14)$$

Typically, for this quantity, we use units of dBZ, so Equation 1.14 becomes Equation 1.15.

$$\hat{Z} = 10 \log_{10} \hat{S} + 20 \log_{10} r + 10 \log_{10} C \quad (1.15)$$

High values of  $\hat{Z}$  are associated with more particles and particles of a larger size in a resolution volume.

#### 1.4.2.3 Doppler Velocity

The oscillation in the discrete I/Q time series in Figure 1.6 is caused by the motion of scatterers relative to the radar. This oscillation is known as the Doppler frequency and is directly proportional to Doppler velocity, given a constant radar wavelength. The time series is sampled at the PRT, so estimates of this frequency are bounded by the Nyquist theorem, which is the cause of Doppler velocity aliasing.

$$v_d \equiv -\frac{f_d \lambda}{2} \quad (1.16)$$

In Equation 1.16,  $v_d$  is the Doppler velocity and  $f_d$  is the Doppler frequency. The frequency can also be estimated using the autocorrelation of the time series, specifically at lag-1. The complex product between a given complex number and the complex conjugate of another has the phase angle equivalent to the difference in phase between the two complex numbers. Thus, the lag-1 autocorrelation of a given I/Q time series yields a complex number whose phase represents the average phase shift across the time series. This phase shift can be used to estimate Doppler velocity using the autocovariance estimator, where  $\hat{v}_d$  is the estimated Doppler velocity,  $v_a$  is the Nyquist velocity given in Equation 1.6, and ‘1’ represents the lag-1 autocorrelation.

$$\hat{v}_d = -\frac{v_a}{\pi} \arg \left( \hat{R}_{XX}(V, 1) \right) \quad (1.17)$$

The Doppler velocity convention is that scatters moving toward the radar have a negative velocity, and those moving away from the radar have a positive velocity. For RaXPol specifically, due to the downconversion chain conventions, the velocities are inverted, and we use the negative of Equation 1.17.

#### 1.4.2.4 Spectrum Width

A measure of the range of velocities present in a resolution volume is known as the spectrum width. A basic estimator of spectrum width  $\sigma_v$  is presented in Equation 1.18. A higher spectrum width might mean turbulence, wind shear, different scatter types, or even noise.

$$\hat{\sigma}_v = \frac{\sqrt{2}v_a}{\pi} \sqrt{\left| \ln \left( \frac{\hat{S}}{|\hat{R}_{XX}(V, 1)|} \right) \right|} \quad (1.18)$$

#### 1.4.2.5 Differential Reflectivity

Polarimetric radars can diagnose additional information about scatterers, especially distinguishing meteorological scatterers from non-meteorological scatterers (Zrnic and Ryzhkov 1999). These radars transmit both horizontally and vertically polarized channels and generate separate I/Q streams ( $V_H$  and  $V_V$ ). The calculations for reflectivity and Doppler velocity in Sections 1.4.2.2 and 1.4.2.3 can be repeated separately for either channel.

$$\hat{R}_{H,V}(l) \equiv \hat{R}_{XX}(V_{H,V}, l) \quad (1.19)$$

We can define special cases of the cross-correlation function to represent the ACF for the horizontal and vertical signal channels, as seen in Equation 1.19. For calculations of reflectivity and velocity with a polarimetric radar,  $\hat{R}_H$  is used by convention.

Figure 1.7 shows the same time series as in Figure 1.6, but only  $I(t)$  for both horizontally and vertically polarized signals. The amplitude of the horizontally polarized signal is greater, partly due to the oblate shape of falling raindrops and partly due to lack of calibration. Oblate raindrops have larger horizontally polarized backscatter cross-sections than vertically polarized ones. Differential Reflectivity ( $Z_{DR}$ ) quantifies the difference in returns with a simple ratio where  $\hat{S}_{H,V}$  are the horizontally and vertically polarized signal power, and  $C$  is a calibration constant resulting from radar biases between horizontal and vertical channels, along with terms from  $C$  in Equation 1.14.

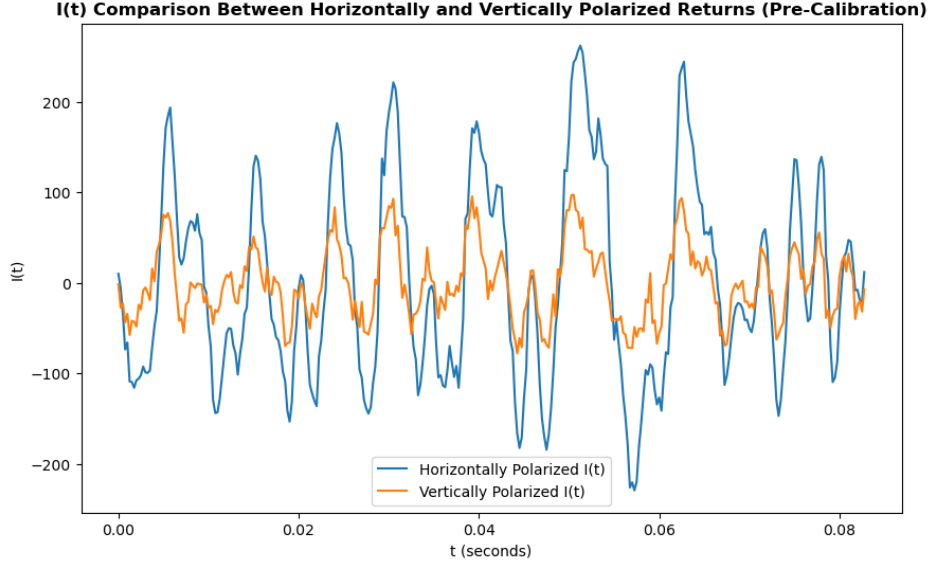


Figure 1.7: Horizontally and vertically polarized time series of  $I(t)$  within the forward flank of a supercell.

$$\hat{Z}_{DR} = 10 \log_{10} \left( \frac{\hat{S}_H}{\hat{S}_V} \right) + 10 \log_{10} C \quad (1.20)$$

A positive  $Z_{DR}$  indicates scatterers wider than they are tall, which include larger raindrops, while a negative  $Z_{DR}$  indicates the opposite.  $Z_{DR}$  close to zero indicates spherical scatterers like small raindrops or hail.

#### 1.4.2.6 Co-polar Cross-Correlation Coefficient

The correlation between equivalent horizontally and vertically polarized I/Q time series is related to the uniformity of scatterers. We can also define a third special case of the cross-correlation function to represent the cross-correlation between  $V_{H,V}$ .

$$\hat{R}_X(l) \equiv \hat{R}(V_H, V_V, l) \quad (1.21)$$

The co-polar cross-correlation coefficient  $\rho_{HV}$ , or simply correlation coefficient, estimates, on average, how the time series vary together. This cross-correlation is scaled by the signal power to get a value between zero and one.

$$\hat{\rho}_{HV} = \frac{|\hat{R}_X(0)|}{\sqrt{\hat{S}_H \hat{S}_V}} \quad (1.22)$$

The correlation coefficient measures scatterer uniformity, so for resolution volumes where scatters are uniform (only rain), values will be very high, whereas in a debris cloud of a tornado, scatterers are non-uniform, and values will be lower.

#### 1.4.2.7 Supercell & Tornado Signatures

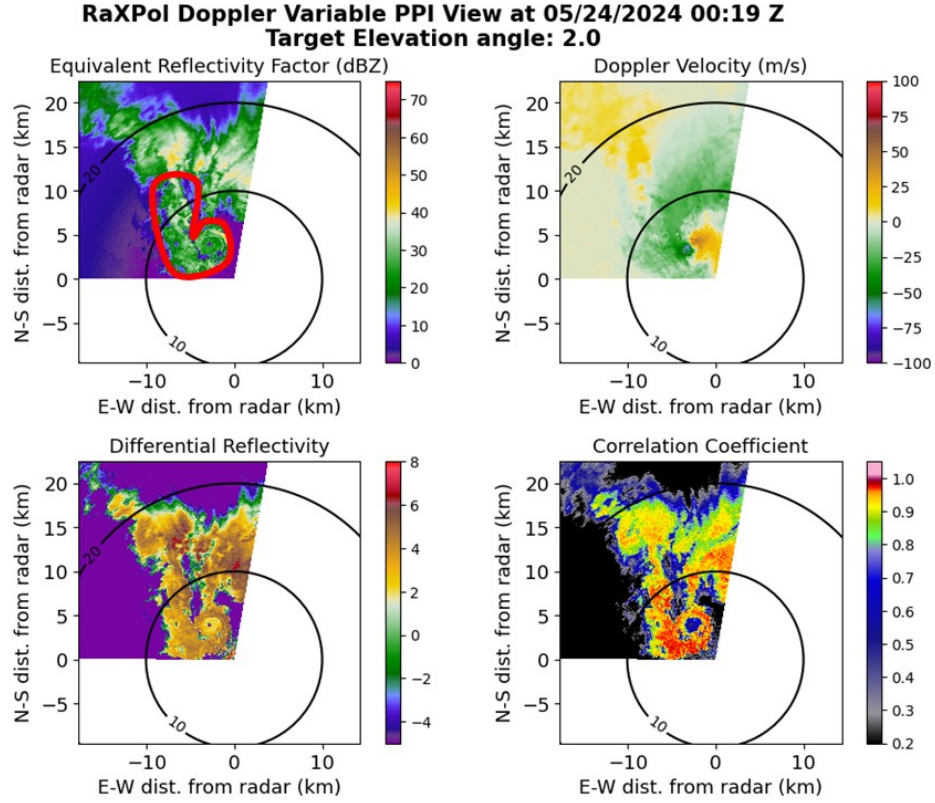


Figure 1.8: RaXPol Doppler variable PPI on May 24, 2024, at 00:19 UTC. The hook echo is circled in red in the reflectivity plot.

Figure 1.8 shows the four radar variables are plotted for a supercell in southwestern Oklahoma on May 23, 2024. In plan position indicator (PPI) view, the hook echo of the supercell is noted, and key features near the hook are described below.

The reflectivity plot in Figure 1.8 depicts a hook echo, a distinguishing feature of supercells as precipitation wraps around the mesocyclone and into the rear-flank downdraft (RFD). At the tip of the hook, a weak echo hole (WEH) is present as an eye-like feature near the center of the tornado vortex (Donaldson 1970). The WEH is caused by the centrifuging of hydrometeors from the tornado cyclone (Dowell et al. 2005). The forward-flank downdraft is visible as the region of precipitation spanning north of the radar. An inflow notch appears as a precipitation-free area around the radar, streaming to the northwest.

In Doppler velocity, this inflow area is also apparent as outbounds to the north of the radar, signifying southerly winds. A tornado vortex signature (TVS) is also present at the tip of the hook echo region, defined by an area of strong inbounds adjacent to an area of strong outbounds in a configuration that suggests cyclonic (counterclockwise in the Northern Hemisphere) rotation (Donaldson 1970).

The differential reflectivity and correlation coefficient plots both show lowered values around the area of the TVS, indicating a tornado debris signature (TDS) (Ryzhkov et al. 2002, 2005). For S-band radars, the TDS is classified by  $\rho_{HV}$  below 0.8,  $Z$  above 45 dBZ, and  $Z_{DR}$  around 0, though some variability exists. The most common aspect of the signature includes low  $\rho_{HV}$ . These values were determined for radars that operate at S-band. Since RaXPol operates at X-band, the variables that rely on signal power ( $Z$  and  $Z_{DR}$ ) can fluctuate even more with Mie scattering regime due to the larger size of debris. Thus, low  $\rho_{HV}$ , for X-band, is the most defining characteristic of a debris cloud.



Polarimetric variables can also change depending on the type of debris lofted. Using numerical simulations, Cross et al. (2023) find that properties of debris like size and dielectric constant can greatly affect debris signatures. For smaller and more reflective debris,  $\rho_{HV}$  can be elevated, lowering the ability for a radar to detect debris. Additionally, Umeyama et al. (2018) found that debris orientation can vary in different parts of the tornado. Distributions of orientation can range from random (near the surface; causing  $Z_{DR}$  near zero), to more vertical (away from tornado center aloft, causing negative  $Z_{DR}$ ), to more horizontal (close to tornado center aloft; causing positive  $Z_{DR}$ ). Furthermore, this study finds that for low  $\rho_{HV}$  characteristic of a debris signature, a mixture of debris and rain needed to be simulated, achieving scatterer heterogeneity.

### 1.4.3 Spectral Analysis

Scatterers will have different velocities in a given resolution volume due to many factors like turbulence, wind shear, or different scatterer sizes and masses (Dowell et al. 2005; Bodine 2014). This results in I/Q time series with multiple Doppler frequencies superimposed. Our primary tool to analyze the distribution of velocity is the power spectral density (PSD). The PSD of a discrete signal is defined as the discrete Fourier transform (DFT) of the cross-correlation function, where  $sS_{X_1, X_2}(f)$  is the signal power spectrum as a function of discrete signal frequency  $f$ .

$$sS_{H,V,X}(f) \equiv \lim_{M \rightarrow \infty} T_s \sum_{l=-(M-1)}^{M-1} R_{H,V,X}(l) e^{-j2\pi f T_s l} \quad (1.23)$$

The first two of these products,  $sS_{H,V}(f)$ , depict horizontally and vertically polarized reflectivity distributions with respect to signal (Doppler) frequency. Since Doppler velocity is a function of Doppler frequency (Equation 1.16),  $sS_{H,V}(f)$  are distributions

of scatterer velocity, with higher distribution values at a given Doppler velocity implying more or larger scatterers traveling at that velocity.

Most meteorological PSDs are Gaussian, but for about 25% of cases, this assumption does not hold (Janssen and Van Der Spek 1985). This is especially true in tornado vortices, where there may be significant shearing or different scatterer types in the resolution volume. With a Gaussian assumption, however, moments of  $sS_{H,V}(f)$  relate directly to estimated quantities computed in Sections 1.4.2.2, 1.4.2.3, 1.4.2.4. The 0th moment relates to reflectivity or the area under the distribution. The 1st moment relates to the distribution mean, which is the estimated Doppler velocity. Finally, the 2nd moment refers to the spectrum width. As such, these processed variables, including polarimetric, are known as moment variables.

Additionally,  $sS_X(f)$  is a cross-spectrum PSD (with the “X” subscript meaning “cross”), containing information about horizontal and vertical channels. It belongs to a class of PSDs known as dual-polarization spectral densities (DPSDs). The two main DPSDs are spectral differential reflectivity ( $sZ_{DR}$ ) and spectral correlation coefficient ( $s\rho_{HV}$ ). These are computed similarly to the bulk differential reflectivity and correlation coefficient variables in Equations 1.20 and 1.22, but instead of bulk statistics of an I/Q time series using the cross-correlation function, we use the full spectra.

$$sZ_{DR}(f) = \frac{sS_H(f)}{sS_V(f)} \quad (1.24)$$

$$s\rho_{HV}(f) = \frac{|sS_X(f)|}{\sqrt{sS_H(f)sS_V(f)}} \quad (1.25)$$

DPSDs are not Gaussian and, instead, they give information about the type of scatterers in a given resolution volume. For example, low values in  $s\rho_{HV}$  at certain Doppler frequencies suggest the presence of non-uniform scatterers, such as tornadic

debris, at those velocities. This is similar to bulk polarimetric variables, as discussed in Section 1.4.2.7.

#### 1.4.3.1 PSD estimation

Since an I/Q time series from any given dwell is not infinite, we must use an estimator.

$$s\hat{S}_{H,V,X}(f) = T_s \sum_{l=-(M-1)}^{M-1} \hat{R}_{H,V,X}(l) e^{-j2\pi f T_s l} \quad (1.26)$$

Note the use of  $\hat{R}_{X_1,X_2}(l)$  in Equation 1.26, showing that the discrete PSD uses the cross-correlation estimator defined in Equation 1.11. This estimator requires the computation of the cross-correlation function before the DFT.

The length of time spent observing a given dwell impacts the PSDs. In Equation 1.26, the number of output points is equivalent to the number of input points. However, the range of frequencies over which the estimator can observe oscillations is constant, bounded by the Nyquist theorem. Thus, the frequency resolution is the ability to distinguish oscillations at nearby frequencies. A high frequency resolution requires a long time series. Related is the concept of a Doppler resolution. Two radars with different carrier frequencies sampling at the same PRT and that collect the same number of discrete time series points will result in frequency spectra with identical frequency resolutions. However, the radar with the lower carrier frequency will have a worse Doppler resolution, as it has a higher Nyquist velocity for a given PRT. Since our project used RaXPol with a constant carrier frequency, we can use the terms ‘frequency resolution’ and ‘Doppler resolution’ interchangeably.

Estimates of PSDs and DPSDs are subject to common limitations of the DFT, which include spectral leakage, bias, and variance. Spectral leakage refers to the pitfall of assuming a signal is infinite. At the edges this signal is not coherent, and this

has the effect of increasing side lobes dramatically, potentially hiding smaller spectral peaks. Bias refers to systematic errors from the estimated spectrum compared to the true spectrum, usually caused by signal processing techniques employed. Variance quantifies the spread of the estimates around the true PSD value due to the randomness inherent in finite data records. Thus, even with the same signal, different segments or realizations can yield significantly different spectral estimates.

A simpler estimator, known as the periodogram estimator, requires the computation of only the DFT of the original I/Q signal.

$$Z_{H,V}(f) = \sum_{m=0}^{M-1} V_{H,V}(m) e^{-j2\pi f T_s m} \quad (1.27)$$

Using the computed DFT, we can obtain estimates of  $sS_{H,V,X}(f)$ .

$$s\hat{S}_{H,V}(f) = |Z_{H,V}(f)| \frac{T_s}{M} \quad (1.28)$$

$$s\hat{S}_X(f) = |Z_H(f)Z_V^*(f)| \frac{T_s}{M} \quad (1.29)$$

For a single dwell, a periodogram by itself has the weakness of very high spectral variance between realizations and high spectral leakage. Methods to correct variance generally involve averaging across the frequency domain, like Daniell's method (Daniell 1946), or averaging multiple "pseudorealizations," or subsets with overlap of the original finite signal, like Welch's method (Welch 1967). Both of these methods come at the cost of reducing frequency resolution in the resultant PSD.

For spectral leakage, a windowing function (e.g., Hanning, Blackman) can be used to limit edge effects and significantly lower sidelobes. However, this can introduce bias by lowering and widening the spectral peak. Furthermore, it can be shown that for a

single dwell periodogram  $s\hat{\rho}_{HV} = 1$ . Thus, some degree of averaging (like in Welch's method) is necessary.

**Bootstrap DPSD** The Bootstrap DPSD method, proposed by Umeyama et al. (2017), aims to preserve frequency resolution while decreasing spectral variance induced by a lack of samples per spectral component in the periodogram method. This is accomplished with a statistical technique known as bootstrapping.

Three instances of the I/Q time series  $V_{H,V}(m)$  of length  $M$  are concatenated, with the left and right concatenations corrected in phase and amplitude by a complex correction factor to ensure coherency across the discontinuity. Then,  $K$  pseudorealizations of length  $M$  are taken from the extended signal in the bootstrapping process. These pseudorealizations are scaled in amplitude such that the average power of all pseudorealizations matches the average power of the original signal. Equation 1.27 is applied to all pseudorealizations to complete  $K$  computations of the DFT for both horizontally and vertically polarized signals. Finally, the  $K$  instances of  $Z_{H,V}$  are averaged and Equations 1.28 and 1.29 are computed. The bootstrapping process also adds bias to DPSD estimates, as samples are not independent. After correcting for this bias and the biases inherent to DPSD estimates outlined in (Yu et al. 2012), the final corrected DPSD estimates,  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ , are obtained.

A few key constraints prevent false extraction of information from where none exists. First, the length of the bootstrapped pseudorealizations cannot exceed the length of the original timeseries. By definition, a Fourier transform cannot provide more frequency resolution than the length of the input time series. Additionally, a longer pseudorealization will have a higher fraction of samples which did not come from the original time series. In order to prevent excessive bias from being introduced, the number of samples which come from the corrected concatenations are limited, which can

be increasingly “warped” the farther they are from the center of the timeseries. These limitations allow this process to preserve, and importantly, not increase frequency resolution while decreasing spectral variance.

This procedure was tested on simulated tornado data by Umeyama et al. (2018), and potential stratification of debris from rain was found. In the simulations, larger debris tended to have non-zero Doppler velocity where wind the circulation should have been perpendicular to the radar. Thus, in the simulation, some centrifuging of heavier debris was observed.

For additional gains in limiting spectral variance, but at the loss of spatial or temporal resolution, multi-dwell averaging can be employed. Range averaging refers to taking multiple I/Q time series from dwells along a radial and averaging  $s\hat{S}_{H,V,X}$  before computation of DPSDs. Similarly, azimuthal averaging can be used. In the temporal dimension, a dwell from a past or future scan can also be used to average, given that the phenomenon being averaged has not moved significantly in the time between scans.

#### 1.4.3.2 Debris Classification

Beyond spectral variance imposed by signal processing methods, DPSDs inside tornadoes may have intrinsically high spectral variance. For bulk  $Z_{DR}$  estimates, Bodine et al. (2014) hypothesized that at low signal-to-noise ratios (SNR) like those in the WEH region of a tornado, a few scatterers may be randomly oriented, yielding high spatial variance. Additionally, Bringi and Chandrasekar (2001) state that low  $\rho_{HV}$  influences the statistics of polarimetric variables as, by definition, horizontal and vertical signals decorrelate. This would also degrade the statistics of polarimetric spectral estimates. Spectral variance in tornado debris signatures is difficult or impossible to avoid, thus becoming a defining characteristic.

Schneider et al. (2025) developed an objective fuzzy logic debris classification algorithm (DCA). Fuzzy logic classification algorithms have been applied to bulk hydrometer classification tasks (Vivekanandan et al. 1999; Straka et al. 2000). In this case, they can be applied to DPSDs to classify spectral components as debris or rain. Spectral input variables for the DCA include  $sZ_{DR}$ ,  $s\rho_{HV}$ ,  $\sigma_{sZ_{DR}}^2$ , and  $\sigma_{s\rho_{HV}}^2$ , where the latter two variables are defined as spectral variance of differential reflectivity and correlation coefficient, and are both comprised of the 9 point moving window variance of the former two variables.

$$A_{class} = \sum_{j=1}^N w_j^{(class)} P^{(class)}(Y_j) \quad (1.30)$$

Equation 1.30 (Equation 1 in Schneider et al. (2025)) defines the aggregation score (between 0 and 1) of a particular class (debris or rain). Here,  $w_j^{(class)}$  is the weight given to a product  $j$  and  $P^{(class)}(Y_j)$  is the membership function result of the product, defined based on idealized simulations run with wood boards as the primary scatterer. The membership function maps the raw input variable to an output between 0 and 1. For any given spectral component, a higher membership for a particular variable for a given class implies that the variable's value increases the likelihood of that spectral component being classified as belonging to that given class.  $A_{class}$  is a spectral variable, as it is an array of values for all Doppler spectral components.

While all input variables are equally weighted for rain,  $\sigma_{sZ_{DR}}^2$  shows high skill in distinguishing debris. Thus,  $A_{debris}$  gives the highest weight to  $\sigma_{sZ_{DR}}^2$  with a weight of 0.4. This leaves  $sZ_{DR}$  with a weight of 0.1, as it shows limited skill in distinguishing debris. For a hard classification of a spectral component, the class with the highest aggregation score is used.

Aggregation parameters can also be used to compute corrected velocity from a PSD.  $A_{rain}$  can be used as fractional measure of rain likelihood. This is accomplished by filtering the spectra by  $W(k) = [A_{rain}]^n$ , where  $n$  is an exponent that determines the aggressiveness of the filter and  $k = fMT_s$  is the spectral component index, with  $n = 8$  being used in the paper. The filter can be used to estimate a velocity by the following equation adapted from Doviak and Zrnic (1993), where  $k_m$  is the spectral component index of maximum power (Schneider et al. 2025).

$$v = -\frac{\lambda}{2M} \left\{ \frac{k_m}{T_s} + \frac{1}{P_H T_s} \sum_{k=k_m-M/2}^{k_m+M/2} (k - k_m) W(k) s S_H [\text{mod}_M k] \right\} \quad (1.31)$$

Using the signal processing techniques described above, we generate a methodology for data collection and data processing to observe low-level structures in tornadoes.



## Chapter 2

### Data & Methods

#### 2.1 Data Collection

Since 2018, a University of Oklahoma-led research effort using RaXPol has been in progress, with the goal of collecting I/Q data within tornado debris clouds for the purpose of computing and analyzing DPSDs, as well as analyzing the feasibility of use in filtering debris. Data collection successes were finally achieved in 2023 and 2024 with varying degrees of success in three different tornadoes through western and central Oklahoma.

##### 2.1.1 Waveform Selection

As in Section 1.2, for a rapid scan rate, a pedestal that can rotate the antenna at  $180^\circ \text{ s}^{-1}$  is used. This has the downside of limiting the number of independent samples in a particular azimuthal dwell at a standard PRT for accurate Doppler velocity estimates. This is remedied by a technique known as “frequency hopping” (Doviak and Zrnic 1993, p. 180), where pulses are sent with varying carrier frequencies to increase the independence of observations, allowing a more accurate sample mean with fewer total samples. Unfortunately, a drawback of this technique is that spectral analysis is not possible, as a frequency-hopped signal becomes incoherent. At the most basic, this

prevents filtering of ground clutter, but additionally, for this project, frequency hopping prevents the spectral analysis of tornado debris clouds. Thus, although frequency hopping provides benefits in allowing rapid scanning, it was not used when the collection of I/Q data was enabled. Additionally, this limitation of a non-frequency-hopped waveform required a slower antenna rotation rate.

As discussed in Section 1.4.3.1, the Nyquist range and frequency resolution of a Fourier transform also depends on two variables, the sampling rate and the length of the time series. The sampling rate is the PRT, while the dish rotation rate controls the length of the time series by increasing the number of samples in a dwell. A higher sampling rate will enable a greater Nyquist range, and a longer time series enables a greater frequency resolution for a constant sampling rate. To evaluate the number of samples needed for reasonable DPSD computation, a slow dish rotation rate of  $6^\circ \text{ s}^{-1}$  was used in most cases. This allowed the computation of spectra with high frequency resolution, with  $M \approx 667 \pm 20$  pulses, uncertainty due to hardware tolerances. By using a slow dish rotation rate, pulses can be subset to evaluate spectra as if using a higher rotation rate and determine an optimal scanning strategy for future data collection.

Since high velocities are involved with tornadic vortices, we must choose a high enough PRT such that velocity aliasing is minimal, as well as accounting for range folding. For RaXPo’s X-band carrier frequency of 9.73 GHz, a PRT of 0.25 ms gives a maximum unambiguous range of 37.5 km and a maximum unambiguous velocity of  $30.8 \text{ m s}^{-1}$ . A higher PRT would result in serious velocity aliasing issues, as tornadoes can easily contain velocities higher than this. A lower PRT would result in issues with range folding, as storms can be large with additional storms nearby, such that significant echoes can originate from outside the maximum unambiguous range. This can contaminate real echoes with erroneous second trip echoes, lowering the quality of data. Thus, a PRT of 0.25 ms was used when storms or tornadoes were in range.

For range measurements, RaXPol typically uses range sampling of 30 meters. Thus, a pulse length of 0.2  $\mu\text{s}$  was used to achieve a range resolution of 30 meters, matching the range sampling.

### 2.1.2 Scanning Strategy

As stated in section 2.1.1, since a slow dish rotation rate was used, conducting full  $360^\circ$  scans scanning would waste time. The area of interest, the tornado, occupies a fairly small volume of space. Thus, sector scanning was implemented, where the radar can slowly scan an azimuthal sector encompassing the tornado. Each successive elevation angle is scanned in the opposite direction, allowing more efficient data collection without the dish having to “reset” to a particular position.

Also important is the range from the radar. With increasing range, angular resolution becomes coarser with increasing distance from the radar, and a resolution volume will include a larger section of the tornado. As a tornado is a small-scale phenomenon, this loss in resolution limits the ability to discern closely spaced areas of different scatterer properties. Thus, as stated by Wakimoto et al. (2012), it is necessary to obtain observations from near a tornado so that small-scale features may be distinguished. Specifically, slow scanning and I/Q data collection was started when a vortex was within 10 km of the radar. With a  $1^\circ$  half-power angular beamwidth, the range at 10 km gives a half-power beamwidth of  $\approx 175$  m.

Finally, target scanning elevation angles also needed to be determined. These varied depending on beam blockage and ground clutter, but the general motivation was to keep the elevation angles as low as possible, attempting to probe the corner flow region and debris cloud. All datasets were collected at different sets of elevation angles, which will be discussed in the next section.

## 2.2 Datasets

### 2.2.1 Cole, OK - May 11, 2023

On May 11, 2023, a dryline located in western Oklahoma initiated multiple storms in an unstable airmass with surface dew points approaching  $70^{\circ}$  F and surface-based convective available potential energy (CAPE) exceeding 2500 J/kg. Wind shear was marginally sufficient to support supercellular convection, with 0-6 km deep-layer shear only around 35 knots (Rasmussen and Blanchard 1998). With a small increase in the low-level jet towards the evening, storms began to produce tornadoes south of the Oklahoma City (OKC) metropolitan area. One of these tornadoes was rated EF-1 and impacted areas in and around Cole, OK (NWS Norman 2023).

This tornado represented the first time RaXPol collected non-frequency-hopped, slow-scan I/Q data through a tornado vortex. It remained within 10 km of RaXPol for much of its life. Unfortunately, for this deployment, the volume coverage pattern (VCP) for sector scanning had not been correctly implemented. This resulted in a large elevation angle drift as the dish translated in azimuth, resulting in a conical scan. Elevation angles in this dataset were as high as  $12^{\circ}$  or even negative, where the radar was primarily scanning the ground. As such, only a single sweep was able to cross the debris cloud, albeit at a high elevation angle, shown in Section 2.3.2.1. Deployment information is given in Table 2.1.

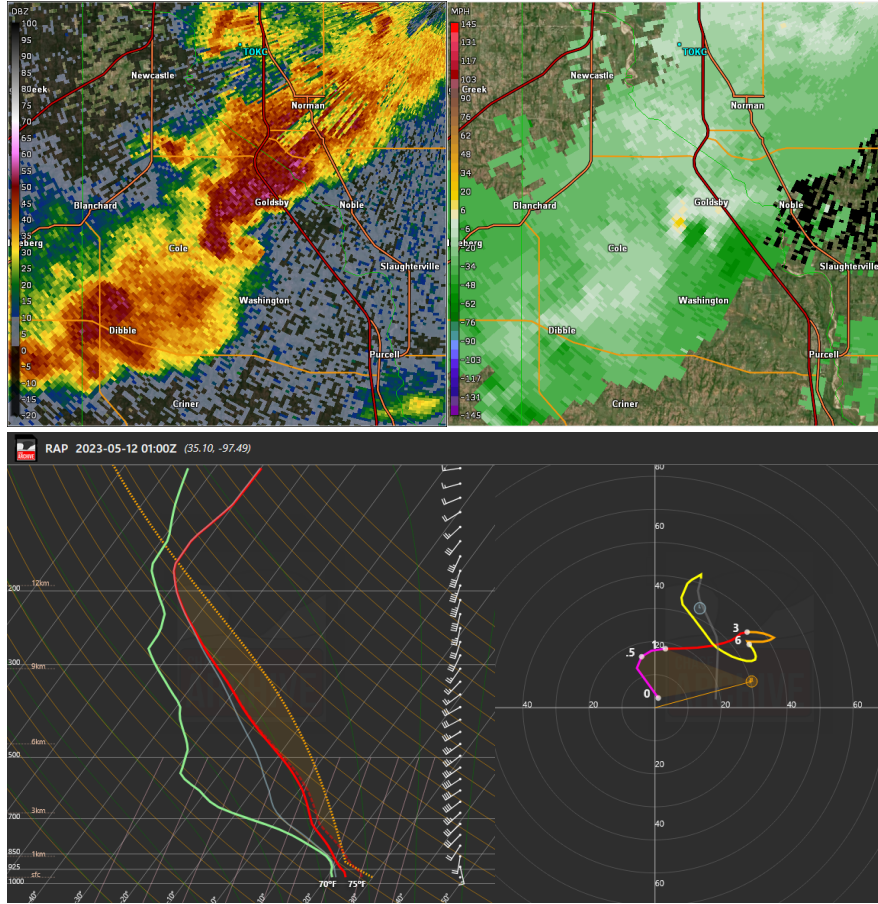


Figure 2.1: NEXRAD KTLX radar reflectivity and velocity PPI view on May 12, 2023, at 01:45 UTC (top). RAP analyzed sounding plotted by Chase Archive (2025) (bottom).

| Cole, OK Deployment Information |                                                      |
|---------------------------------|------------------------------------------------------|
| Location                        | 35.1823° N, 97.5272° W                               |
| I/Q Data Collection Time        | May 12, 2024 - 01:40 to 02:02 UTC                    |
| Pulse Repetition Time           | 0.25 ms                                              |
| Nyquist Velocity                | 30.8 m s <sup>-1</sup>                               |
| Pulse Width                     | 30 m                                                 |
| Elevation Angles                | Variable - Negative to 12°                           |
| Dish Rotation Rate              | Variable - 6° s <sup>-1</sup> to 12° s <sup>-1</sup> |

Table 2.1: Deployment information during I/Q data collection for the Cole, OK tornado.

### 2.2.2 Union City, OK - May 19, 2024

Following an update in the sector scanning VCP in mid-May 2024, a long stretch of severe weather looked to unfold over the Southern Plains. On May 19, a dryline in the Eastern Texas panhandle initiated a complex of storms that would cross into Western Oklahoma. Ahead of these storms lay surface-based instability exceeding 2500 J/kg with dew points around 65° F. Deep-layer shear would support supercells, with 0-6 km shear exceeding 35 knots. Out of these storms, a dominant supercell formed, moving east-southeast towards the OKC metro. This supercell spawned an EF-2 tornado near Custer City, OK, along with a few more weak tornadoes as it propagated to the southeast. After nightfall, the supercell produced two simultaneous tornadoes, one rated EF-2 and one rated EF-0, northeast of Union City, OK (NWS Norman 2024a).

On this day, there were new challenges to data collection. Towards the middle of field operations, the vertical channel stopped delivering normal returns, meaning that the computation of polarimetric variables was impossible. However, towards the end of the day, the vertical channel partially recovered, and some polarimetric variables could be computed with potentially unknown biases. During the final two tornadoes, after the vertical channel had partially recovered, RaXPol was west of the storm with tornadoes beyond 25 km, farther than the 10 km desirable radius. Despite these problems, these tornadoes presented a good opportunity to test the new sector-scanning VCP. Thus, the team collected slow-scan I/Q data of the distant tornadoes. Additionally, instead of trying different dish rotation rates, the team settled on  $6^\circ \text{ s}^{-1}$ . With a very slow rate, faster rates of rotation can be simulated by subsetting the time series, giving fewer pulses in a dwell. Deployment information is given in Table 2.2.

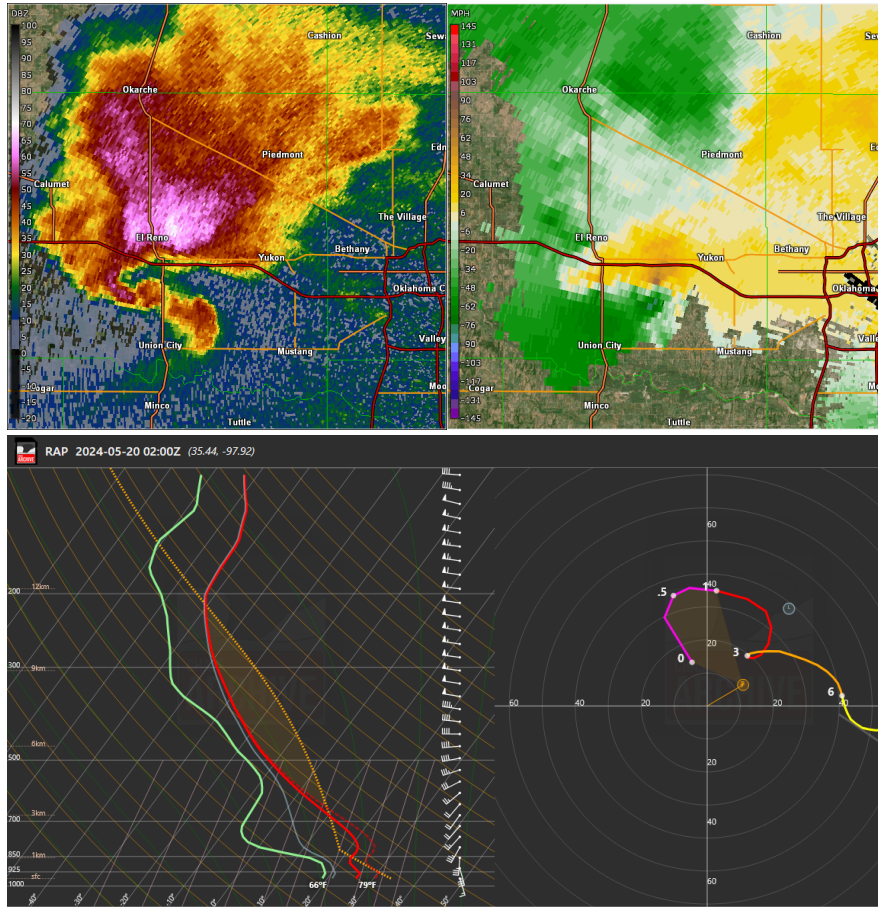


Figure 2.2: NEXRAD KTLX radar reflectivity and velocity PPI view on May 20, 2024, at 02:30 UTC (top). RAP analyzed sounding plotted by Chase Archive (2025) (bottom).

| Union City, OK Deployment Information |                                   |
|---------------------------------------|-----------------------------------|
| Location                              | 35.4079° N, 98.1843° W            |
| I/Q Data Collection Time              | May 20, 2024 - 02:38 to 02:56 UTC |
| Pulse Repetition Time                 | 0.25 ms                           |
| Nyquist Velocity                      | 30.8 m s <sup>-1</sup>            |
| Pulse Width                           | 30 m                              |
| Elevation Angles                      | 0°, 1°, 2°, 3°, 4°, 5°            |
| Dish Rotation Rate                    | 6° s <sup>-1</sup>                |

Table 2.2: Deployment information during I/Q data collection for the Union City, OK tornado.

### 2.2.3 Eldorado, OK - May 23, 2024

On May 23, 2024, a dryline appeared once again over the Eastern Texas Panhandle. A strongly unstable airmass lay east of the dryline, with dew points approaching the mid-70s (°F), and surface-based CAPE exceeding 4000 J/kg. Deep layer shear was once again favorable for supercells, with 0-6 km shear around 40 knots. Storms initiated in the far Eastern Texas Panhandle just before 3:00 pm central daylight time (CDT) and moved into Western Oklahoma, quickly becoming supercellular. A cell that initiated near Childress, TX, would become dominant, producing a significant (EF-2) tornado north of Eldorado, OK, after crossing the state line (NWS Norman 2024b).

During this day of field operations, a similar problem affected the vertical channel during early deployments while storms were initiating and organizing. Similar to May 19, the vertical channel recovered towards the final deployment of the day. Data collection began at the intersection of OK state highways 34 (north/south) and 6 (east/west), located south of East Duke and north of Eldorado, Oklahoma. During this deployment, RaXPol collected high-quality I/Q data through the tornado vortex, but potentially with unknown biases in polarimetric variables.

Initially, the mesocyclone was located outside the 10 km optimum radius for I/Q data collection. RaXPol collected volume scans in rapid-scan mode, documenting the tornadogenesis process. After tornadogenesis, the tornado moved to the southeast towards the deployment location and exhibited multi-vortex characteristics. When the tornado was within 10 km of the radar, the scanning strategy was changed to slow-scanning I/Q data collection. This strategy continued as the tornado made a left turn, avoiding the deployment location but coming within 4 km of RaXPol. The 0° through 5° elevation angles were scanned, albeit with three repeated 0° in a volume to gain additional temporal resolution at the lowest levels (see the “Elevation Angles”



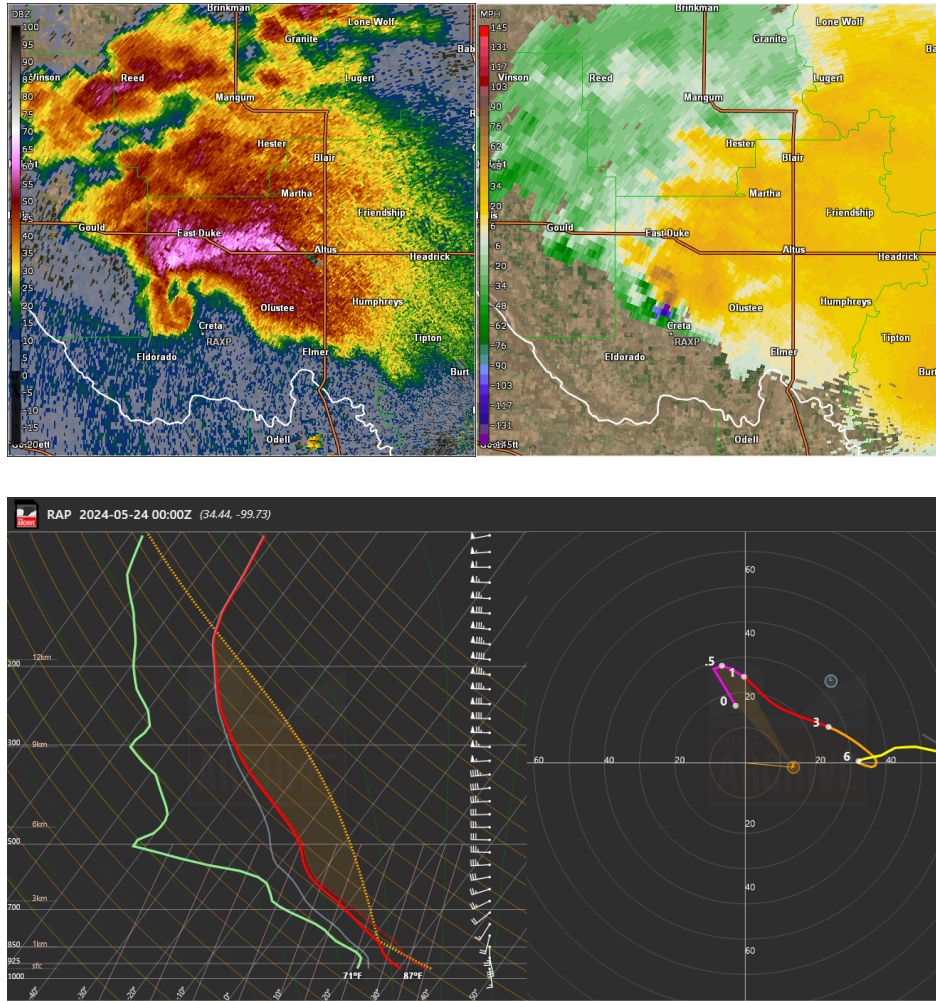


Figure 2.3: NEXRAD KFDR radar reflectivity and velocity PPI view on May 24, 2024, at 00:18 UTC (top). RAP analyzed sounding plotted by Chase Archive (2025) (bottom).

row in Table 2.3). As the tornado passed to the north and crossed OK-34, the sector and elevation angles were adjusted to account for the tornado moving out of the new sector and behind a hill, which caused beam blockage. Thus, 2° of elevation was added to all sweeps in the volume. Also, after passing to the north of RaXPoI, a weak satellite (EF-0) was observed, merging with the main vortex, which reformed as a large stovepipe-shaped tornado. It continued to drift to the northeast into the forward-flank precipitation before dissipating. Over 40 minutes of I/Q data were collected during

this deployment. Because this is the only dataset with significant quantities of I/Q data collected at predictable elevation angles through the debris cloud of a tornado, this was the primary dataset used for spectral analysis. Deployment information is given in Table 2.3, and a deployment map is shown in Figure 2.4.

| Eldorado, OK Deployment Information |                                                      |
|-------------------------------------|------------------------------------------------------|
| Location                            | 34.51249° N, 99.56151° W                             |
| I/Q Data Collection Time            | May 24, 2024 - 00:08 to 00:52 UTC                    |
| Pulse Repetition Time               | 0.25 ms                                              |
| Nyquist Velocity                    | 30.8 m s <sup>-1</sup>                               |
| Pulse Width                         | 30 m                                                 |
| Elevation Angles                    | 0°, 0°, 0°, 1°, 2°, 3°, 4°, 5° (all +2° behind hill) |
| Dish Rotation Rate                  | 6° s <sup>-1</sup>                                   |

Table 2.3: Deployment information during I/Q data collection for the Eldorado, OK tornado.

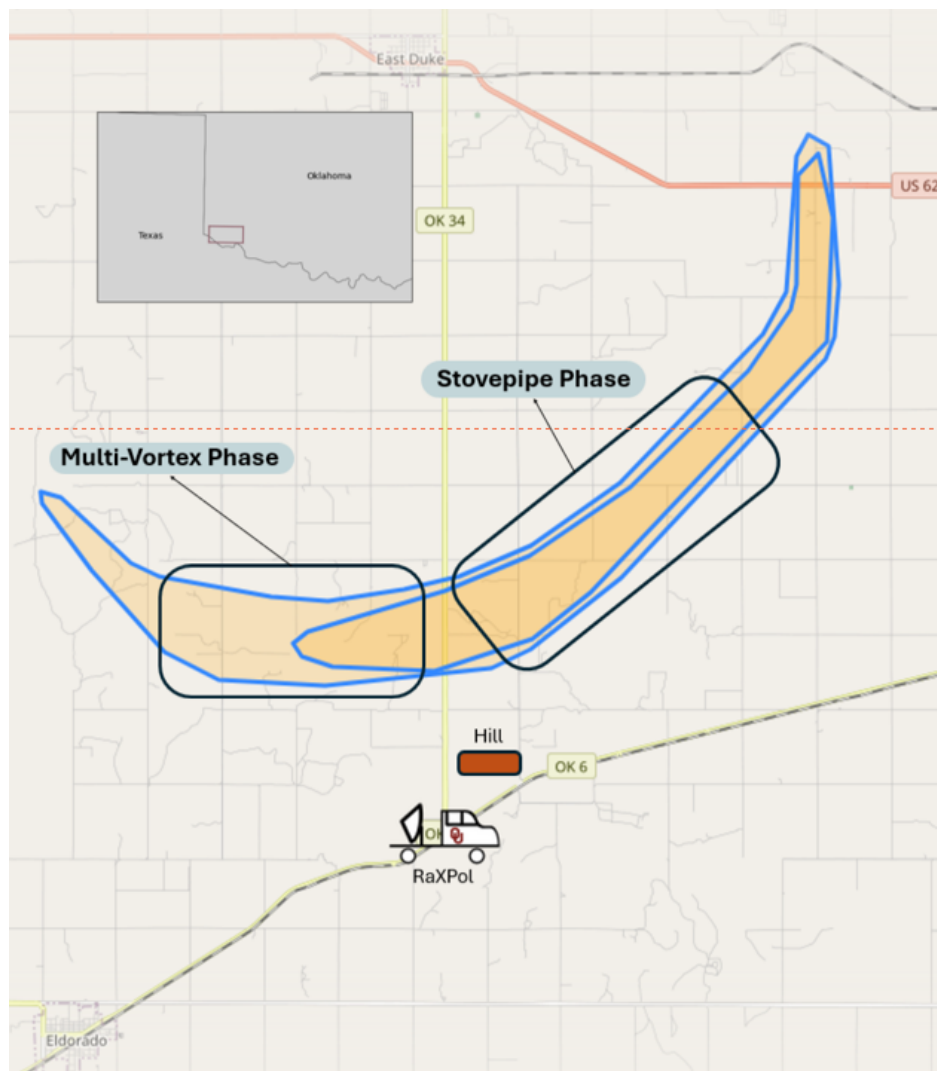


Figure 2.4: Map showing deployment location from Table 2.3. Tornado damage intensity contours from NOAA (2024) surveyed by NWS Norman.

## 2.3 Data Processing

### 2.3.1 Moment Variable Calculation

Calculation of moment variables which are plotted in PPI view, discussed in Sections 1.4.2.2 to 1.4.2.7, requires taking multiple time series of I/Q data from a given dwell, computing mean statistics, and plotting as a “pixel”. This process of grouping pulses into dwells and calculating statistics is repeated for all data. The azimuthal width of adjacent pixels is known as the azimuth spacing interval. By convention, the azimuth dwell length and azimuth spacing interval of dwells matches the half-power beamwidth of the radar, as all pulses at a given range will have resolution volumes including the center of the dwell. This is known as standard resolution processing.

Additional methods of processing are also possible. Due to the coarse nature of radar PPI grids, velocities in tornadoes can be underestimated, as data in areas with lower velocity are averaged with data in areas of higher velocity, especially with range. Super resolution processing involves grouping I/Q data into dwells  $0.5^\circ$  wide with a  $0.5^\circ$  azimuth spacing interval, doubling the resolution in the azimuthal direction (Brown et al. 2002). This increased resolution allows for the detection of tornado vortices at longer ranges. More recently, Nai et al. (2020) showed that even narrower azimuth spacing intervals, up to  $0.125^\circ$ , can decrease velocity underestimates further.

For spectral analysis, these techniques have the limitation of decreasing dwell length. A compromise can be struck between dwell length and azimuth spacing interval with a technique known as azimuthal oversampling (Melnikov et al. 2015). Longer dwell lengths are used with a shorter azimuth spacing interval, such that some I/Q data points are used multiple times in a given PPI grid. This method also improves the detection of vortices, as the azimuthal resolution is similarly increased. The added

benefit, however, is that a bulk value of a radar variable displayed on a PPI plot corresponds to spectra with a high frequency resolution. This is important with spectral analysis, as we expect bulk variables to exhibit matching statistics of the spectra that were generated from the data that was used to calculate them. Thus, for the processed moment data, an azimuth spacing interval of  $0.5^\circ$  was used with a dwell length of  $1^\circ$ , matching RaXPol’s half-power beamwidth.

For the data pipeline, I/Q data were processed into CFRadial format (Dixon and Lee 2016) using a custom Python script that allows azimuth spacing intervals and dwell lengths. From CFRadial, the Python Atmospheric Radiation Measurement Radar Toolkit (Py-ART) library (Helmus and Collis 2016) was used to plot and manipulate radar data. The Lidar Radar Open Software Environment (LROSE) toolkit (DeHart et al. 2024) was used to convert between CFRadial and DORADE formats with Radx-Convert. DORADE format was used with solo3 to de-alias velocity data.

### 2.3.2 PSD & DPSD Calculation

For the calculation of PSDs and DPSDs, as discussed in Section 1.4.3.1, the Bootstrap DPSD (Umeyama et al. 2017) was used. The process of computing PSDs and DPSDs, regardless of method, requires several parameters: the discrete I/Q time series ( $V(m)$ ), a data windowing function, an SNR cutoff, averaging in frequency space (similar to Daniell’s method), the number of dwells to average in either a radial, azimuthal, or temporal direction, and finally, the number of fast-Fourier transform (FFT) input and output points. The Bootstrap DPSD method requires one additional input: the number of pseudorealizations of the extended time series to average. After the computation of spectra, if there were aliasing, spectra were de-aliased manually using custom python software.

**Data Windowing** For the data windowing function, a Blackman window was used, but other windows that were tested, like Hann, Hamming, and Blackman-Harris, yielded very similar spectra. A rectangular window resulted in issues with sidelobe propagation for DPSD estimates ( $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ ) and erroneously high values for parts of the spectra where SNR was very low, which suggested significant spectral leakage. Thus, a rectangular window was not suitable for this analysis.

**SNR Cutoff** Unlike PSDs, DPSDs can contain data which appears legitimate even at very low SNRs. These data do not have associated reflectivity returns and are likely to be mainly noise. Schneider et al. (2025) notes that spectral variances for returns with a low SNR are usually high and can lead to a misclassification of debris in areas of spectra where SNR is low. Thus, all spectra had an SNR cutoff. Values where horizontally polarized SNR was below 20 dB were masked, although in some instances, this cutoff had to be lowered because of weaker returns from low scatterer counts overall. Horizontally polarized SNR (SNRH) was chosen specifically due to the issue with the vertical channel during the Eldorado, OK deployment, where the noise measurement from the vertical channel seemed to be anomalously high despite normal-appearing returns. To avoid masking too much signal, only SNRH was used, and this convention was retroactively applied to the other two cases.

**Frequency Space Averaging** Due to high frequency resolution and high spectral variance in spectra through the tornado vortex, frequency space averaging was used to generate human-interpretable spectra for line plots, similar to Daniell’s method. This was done with a Savitzky–Golay filter using a 2nd-degree polynomial fit. A reasonable estimate of the discrete spectrum width is determined by the number of points in a PSD that have been masked by SNR. The filtering window used was equivalent to 10% of the SNR-masked spectral points. An example PSD line plot with and without filtering

is shown in Figure 2.5. Spectra were not filtered in frequency space when displayed in batch with range-Doppler plots (see Figure 2.7) or before use with the DCA.

**Dwell Averaging** No radial, azimuthal, or temporal averaging was used in the computation of PSDs or DPSDs. Spatially, sharp velocity gradients in the tornado were shown to manifest spurious multimodal distributions in PSDs, potentially leading to misinterpretation of spectra as bimodal owing to different scatterer motions in a resolution volume. Averaging in space also resulted in smoothing of important peaks in PSDs. An example of this is seen in Figure 2.6. Temporally, by slow scanning, the tornado statistics were not approximately stationary with each sweep. Thus, averaging in this dimension would yield tornado spectra averaged with regular storm spectra or clear air spectra.

**Input FFT Points** For spectra that were analyzed qualitatively and used for the fuzzy DCA, the number of input FFT points was the length of the full  $1^\circ$  azimuthal dwell. An additional objective was to synthetically test the quality of spectra at different dwell lengths. At a faster dish rotation rate for an equivalent azimuthal dwell length, the number of points in a discrete I/Q time series will be lowered. This can be emulated by subsetting the I/Q time series by some points less than its full length. In our case, the original un-subsetted time series has an azimuthal dwell length equal to the half-power beamwidth ( $1^\circ$ ). The effective beamwidth, defined in Equation 7.34 of Doviak and Zrnic (1993, p. 196), of a subsetted time series would be within  $\approx 0.45^\circ$  of an analogous time series with an increased rotation rate and decreased total number of points such that the azimuthal dwell length is equivalent to  $1^\circ$ . As such, we

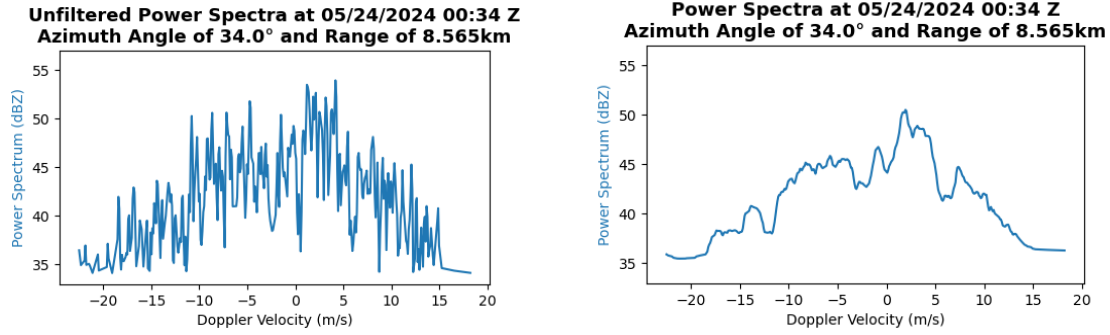


Figure 2.5: PSD for a dwell collected from RaXPol before and after filtering. Filtering improves interpretability by reducing high frequency variations, revealing underlying patterns of the data at the cost of frequency resolution. Low-pass filtering was not used in range-Doppler plots like Figure 2.7, as it was not needed for interpretability.

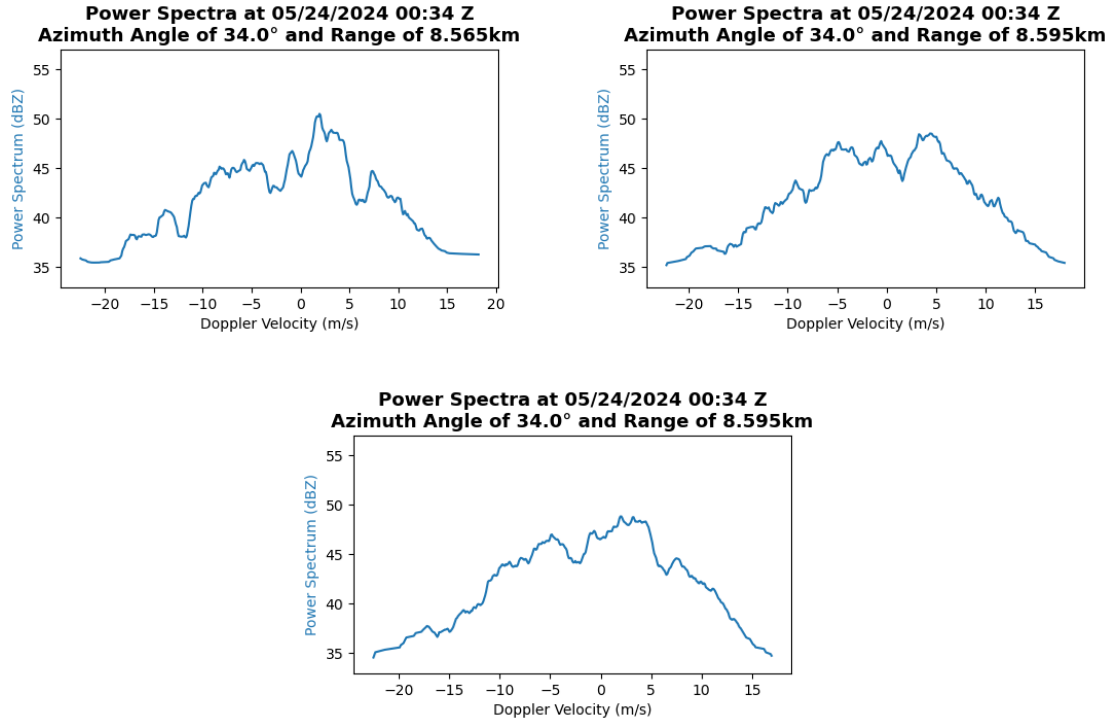


Figure 2.6: Top row) Two adjacent single-dwell PSDs. The PSD at 8.565 km has a distinct peak and a sub-peak, while the PSD at 8.595 km has a wider peak. Spectral characteristics exhibit significant spatial variability, as demonstrated in the comparison between these two adjacent dwells. Bottom row) The range averaged spectrum. Some resolution is lost, like the ability to distinguish the main peak in the top left spectrum.



can emulate reduced frequency resolution while maintaining a similar effective resolution volume (the resolution volume of a dwell using effective beamwidth) through subsetting.

**Output FFT Points** For spectra that were either analyzed qualitatively and or used as inputs for DCA spectral variables, the number of output FFT points equaled the length of the time series input into the FFT. However, when testing spectral quality across different pseudorealizations and dwell lengths, the output of the Bootstrap DPSD algorithm needed to include the same number of spectral components regardless of the number of input pulses. Thus, the power of two larger than the length of the original dwell prior to subsetting was used as the number of output FFT points when qualitatively testing the quality of spectra. Since  $M \approx 667 \pm 20$  pulses, 1024 output points were used in the spectral component convergence algorithm.

**Number of Pseudorealizations** An increased number of pseudorealizations can decrease spectral variance from signal processing with the only cost being computational complexity. Umeyama et al. (2017) used 20 pseudorealizations in their analysis, but it is possible that using more would result in improved spectra. The effect of increasing pseudorealizations will be tested.

### 2.3.2.1 Example Spectra From Earlier Datasets

We can calculate some example spectra from earlier datasets before moving on to the Eldorado, OK dataset. The Cole, OK dataset represented the first time RaXPol had successfully collected non-frequency-hopped I/Q data through a tornado vortex, potentially sampling debris.

**DPSD Range-Doppler Plots** PPI plots and spectra plots are given in Figures 2.7 and 2.8. The pink lines in the PPI plots denote dwells along which spectra were computed in batch. The bottom 4 plots show the spectra computed via Bootstrap DPSD in range-Doppler plots at 1000 pseudorealizations, with the X-axis as spectral frequency components converted to Doppler velocity, and the Y-axis as spectra computed from individual dwells. This PPI and range-Doppler plot style is used frequently through this work.

Figure 2.7 shows a visualization of spectra along a radial through the inbound winds of the tornado vortex. In this particular sweep, the dish rotation rate was  $12^\circ \text{ s}^{-1}$ , giving  $\approx 333$  points in a dwell. Spectral components with high SNR (unmasked) show an increase in Doppler velocity towards the center of the vortex (7.4 km). This tornado generally had weaker reflectivities, so a lower SNR cutoff of 10 dB was used. The power spectra show the WEH, as values near peak velocities are lower and almost fully masked by the SNR cutoff. Areas of potential debris are visible as lower  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ .

Beam height above radar level along the radial was calculated by using the 4/3 earth model (Doviak and Zrnic 1993). With the small angle approximation, we add the pedestal height of 2.5 meters to the beam height to get the beam height above the ground level of the radar. To account for varying topography, the difference between the elevation of the radar location and the dwell location was also accounted for using the USGS National Elevation Dataset at 10-meter resolution (USGS 2020), made available by the Open Topo Data Application Programming Interface (API).

In Figure 2.8, we observe similar spectra to those in Figure 2.7. Velocity increases and  $s\hat{S}_H$  decreases towards the vortex center (25.5 km). During this deployment, a dish rotation rate of  $6^\circ \text{ s}^{-1}$  was used, giving double the frequency resolution. Also,

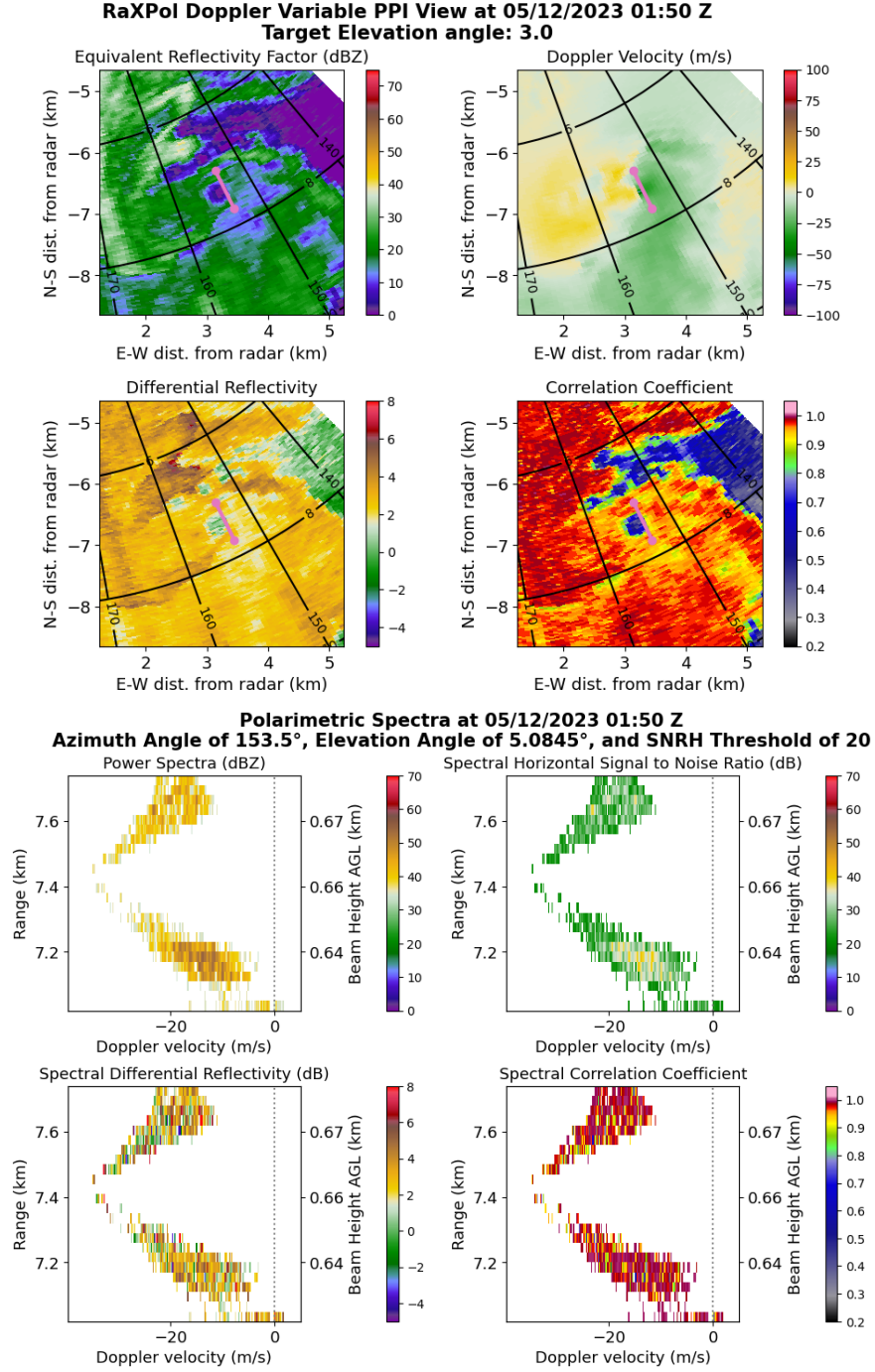


Figure 2.7: RaXPol Doppler variable PPI view at 01:50 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Cole, OK EF-1 tornado.

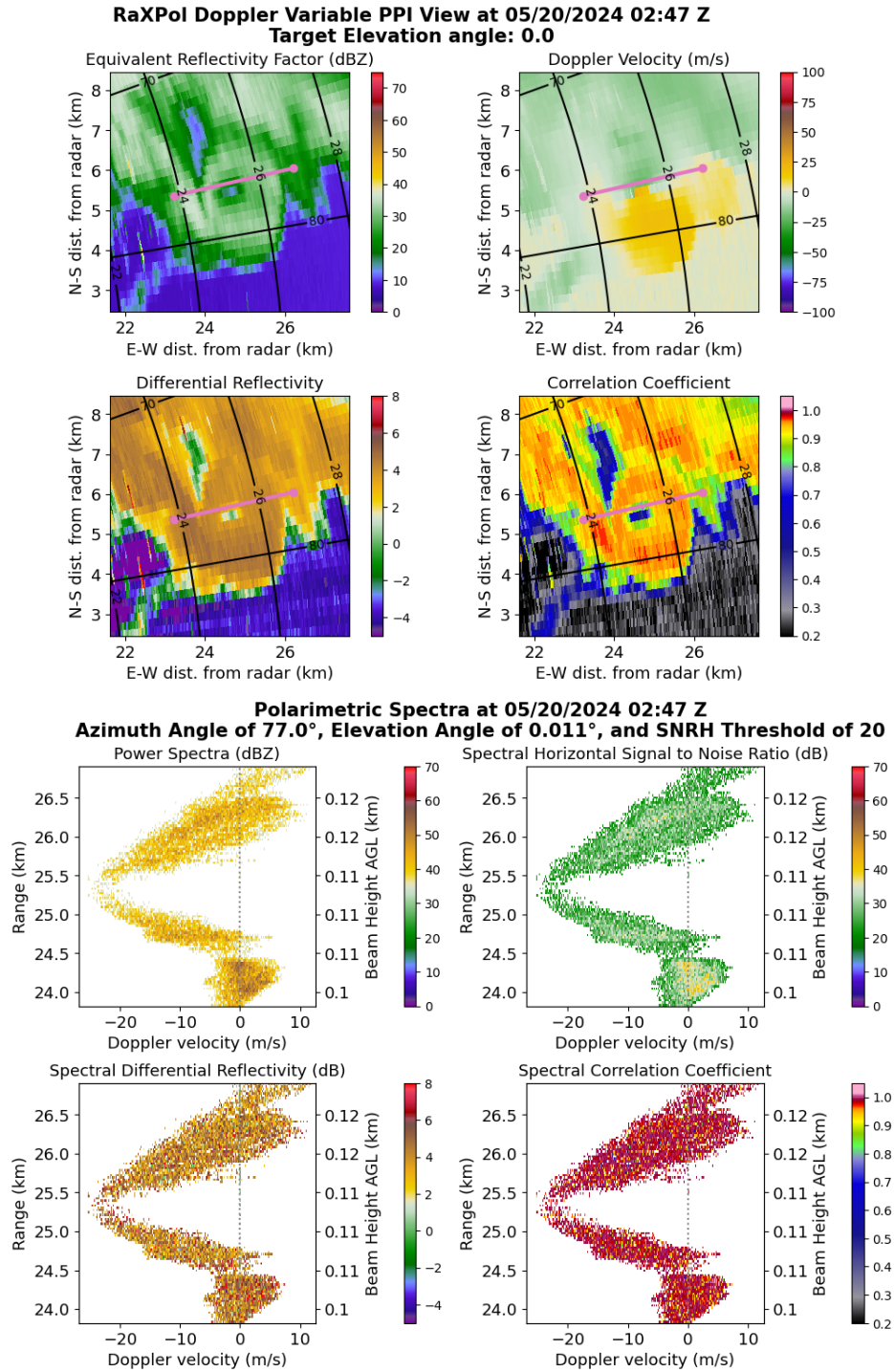


Figure 2.8: RaXPoI Doppler variable PPI view at 02:47 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Union City, OK EF-0 tornado.

unlike in Figure 2.7, we do not observe lower values of  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$  near the vortex center, potentially indicating limited to no debris in the resolution volumes. Spectra from the Eldorado, OK tornado will be discussed in Chapter 3.

### 2.3.2.2 Fuzzy DCA Calculation

The calculation of variables associated with the debris classification algorithm (DCA) outlined in Schneider et al. (2025) was conducted identically to the methods described in the paper, with no change to the membership functions or changes to the calculation of spectral variances (a 9-point moving window is used). When filtering  $s\hat{S}_H$ , the  $[A_{rain}]^n$  exponent was kept at  $n = 8$ , and corrected velocities were computed as well. To mitigate the low SNR misclassification issue mentioned in the paper, only the spectral components above the SNR cutoff were used in computations involving the DCA. In addition, a normalized rain aggregation parameter  $An_{rain}$ , which is defined in Equation 2.1, was computed. This field is greater than 0.5 when a spectral component will be classified as rain.

$$An_{rain} = \frac{A_{rain}}{A_{rain} + A_{debris}} \quad (2.1)$$

**DCA Variable Range-Doppler Plots** DCA variable spectra are computed and displayed in range-Doppler plots, examples of which are shown in Figures 2.9 and 2.10. The top row includes 9-point window spectral variances of DPSD fields ( $\sigma_{sZ_{DR}}^2$  and  $\sigma_{s\rho_{HV}}^2$ ). The lower left is  $An_{rain}$ , where red-colored spectral components are likely to be rain and blue are likely to be debris. The lower right is the horizontally polarized PSD ( $s\hat{S}_H$ ) filtered by  $[A_{rain}]^8$ , which is used to estimate corrected velocity. This plot also contains autocovariance estimated velocity ( $V_{AC}$ , red line) and velocity estimated

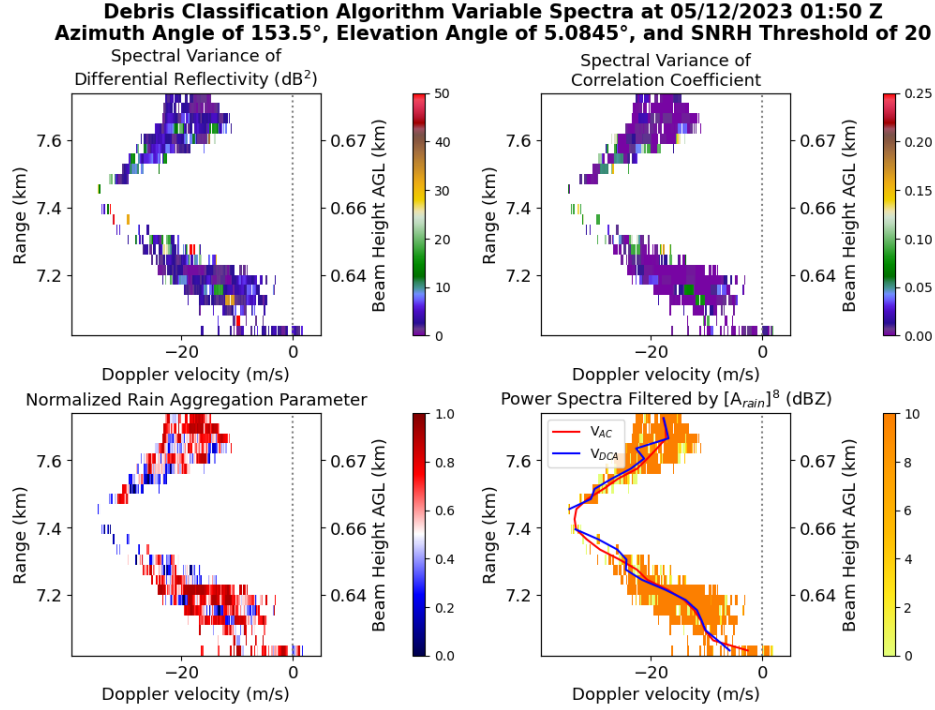


Figure 2.9: DCA (Debris Classification Algorithm) variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 2.7 for the Cole, OK EF-1 tornado.

by the filtered  $s\hat{S}_H$  by  $[A_{rain}]^8$  ( $V_{DCA}$ , blue line), computed with Equation 1.31. This aims to show how the filter affects velocity estimates. Each DCA variable range-Doppler plot is tied to an associated DPSD range-Doppler plot, which is referenced in the caption.

In Figure 2.9, the DCA flagged some spectral components as debris. Specifically in the more distant ranges, greater than 7.4 km, the slower velocities are flagged as debris, allowing  $V_{DCA}$  to be slightly greater than or equal to  $V_{AC}$  for many dwells. For the dwells closer to the radar where a significant portion of the signal is still retained (closer than 7.3 km), this pattern persists  $V_{DCA}$  and  $V_{AC}$ . For dwells where all or the majority of spectral components are lower than the SNR cutoff, there is no estimated  $V_{DCA}$ , or  $V_{DCA}$  estimates are of low quality.

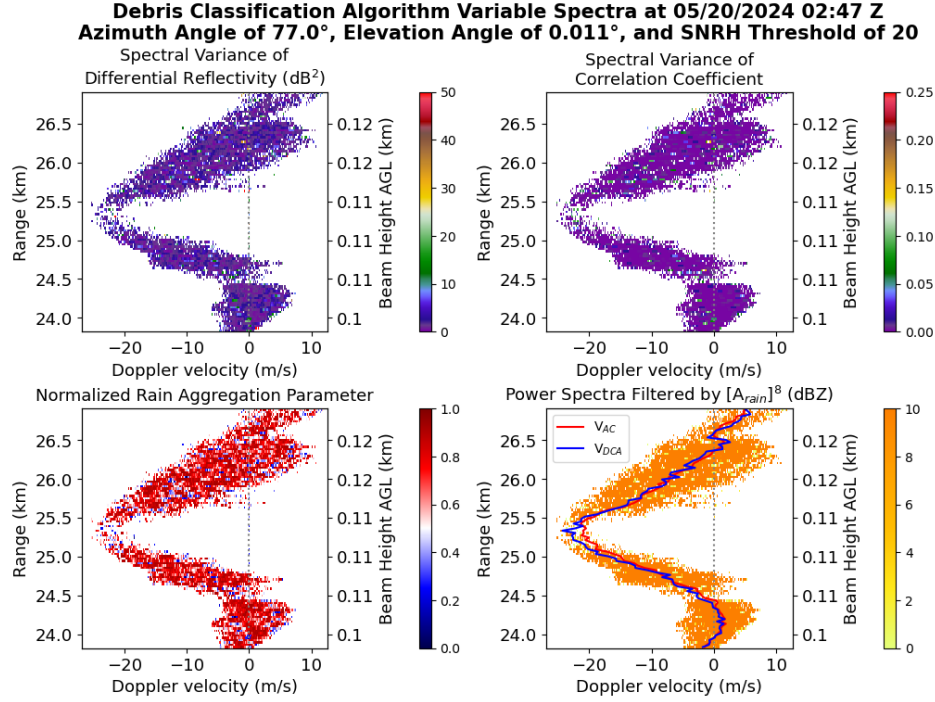


Figure 2.10: DCA (Debris Classification Algorithm) variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 2.8 for the Union City, OK EF-0 tornado.

In Figure 2.10, the spectral components flagged by the DCA as debris appear randomly distributed, likely owing to noise. At the peak of the vortex just under 25.5 km, DCA-estimated velocities are slightly stronger than autocovariance-estimated velocities.

Overall, while not necessarily valuable in diagnosing areas of debris or rain, the Cole, OK and Union City, OK cases prove useful in confirming that the process of DPSD and DCA variable estimation provides output that is physical. All plots show the increase in wind magnitude closer to the tornado center. Even though the polarimetric characteristics are less insightful due to distance from the radar and beam height, the ability of DPSD and DCA variables to identify spectral components as rain instead of noise is encouraging. As in Section 2.3.2.1, DCA products for the Eldorado, OK case will be discussed in Chapter 3.

### 2.3.3 Spectral Component Convergence Algorithm

It is important to assess is the ability of faster dish rotation rates to produce quality spectra representative of the scatterer velocity and type distributions in the effective resolution volume. In order to assess this quantitatively such that we can be confident there is limited bias and limited variance in spectra, we define a convergence metric for spectral components.

At a fixed number of pseudorealizations, we assume that with increasing frequency resolution in our spectra, the value of a given spectral component should approach a “true” value. Thus, we can test for convergence of spectral components to a value as the number of input points into the Bootstrap DPSD algorithm increases. We can keep track of spectral component change as the absolute value of the difference between spectral components for spectra generated by an input time series of length one longer than the previous, like the following, where  $s_{k,N}$  is the value of a specific spectral component  $k$  in any given spectrum  $s$ , where the number of input points in the time series to generate that spectrum was  $N$ .

$$\Delta s_{k,N} = |s_{k,N} - s_{k,N-1}| \quad (2.2)$$

With a limited number of pulses, changes in spectral components are assumed to be large, as frequency resolution increases rapidly. Thus, we take the maximum values of  $\Delta s_{k,N}$  for  $N$  ranging from 3 to 10 ( $\max(\Delta s_{k,N[3,\dots,10]})$ ). We can say a spectral component has converged when for three consecutive values of  $N$  where  $N$  is greater than 10,  $\Delta s_{k,N[11,\dots]}$  is less than 10% of the maximum of the values for  $\max(\Delta s_{k,N[3,\dots,10]})$ . If three consecutive values are never less than 10% of  $\max(\Delta s_{k,N[3,\dots,10]})$  by the time



| Debris Dwells Used in Convergence Algorithm |                 |               |                  |
|---------------------------------------------|-----------------|---------------|------------------|
| Sweep Time                                  | Elevation Angle | Azimuth Angle | Ranges (km)      |
| 00:19 UTC                                   | 0°              | 317.5°        | 4.575 to 5.145   |
| 00:20 UTC                                   | 4°              | 333.5°        | 4.095 to 4.665   |
| 00:29 UTC                                   | 3°              | 26.5°         | 6.495 to 7.065   |
| 00:11 UTC                                   | 0°              | 339.5°        | 4.035 to 4.605   |
| 00:39 UTC                                   | 2°              | 32.5°         | 11.175 to 11.745 |

Table 2.4: Selected debris dwells for use in convergence algorithm. Each row contains 20 dwells. Dwells were picked from a variety of tornado stages and elevation angles where there were significant low  $\rho_{HV}$  and  $Z_{DR}$  returns. All dwells were taken from near or in the tornado cyclone, as this is the only location where debris can be found.

| Rain Dwells Used in Convergence Algorithm |                 |               |                  |
|-------------------------------------------|-----------------|---------------|------------------|
| Sweep Time                                | Elevation Angle | Azimuth Angle | Ranges (km)      |
| 00:19 UTC                                 | 2°              | 311.0°        | 4.695 to 5.265   |
| 00:09 UTC                                 | 4°              | 316.0°        | 8.205 to 8.775   |
| 00:18 UTC                                 | 5°              | 350.5°        | 10.995 to 11.565 |
| 00:21 UTC                                 | 0°              | 331.0°        | 4.005 to 4.575   |
| 00:33 UTC                                 | 7°              | 25.5°         | 8.625 to 9.195   |

Table 2.5: Selected rain dwells for use in convergence algorithm. Each row contains 20 dwells. High  $\rho_{HV}$  and  $Z_{DR}$  dwells were picked from primarily near-tornado locations, but in the 00:18 UTC sweep, dwells were picked from within the forward-flank downdraft region. The dwells from the 00:19 UTC and 00:21 UTC sweeps were near enough to the tornado to contain aliased velocities.

$N = M$ , that spectral component is unconverged. The unconverged percent of spectral components can give a qualitative insight into the quality of spectra.

For the Eldorado, OK case, the only one with significant debris, this algorithm was applied to 100 observed debris dwells (within a TDS) and 100 rain dwells, and differences in convergence statistics were observed. These dwells were taken from a variety of sweeps, with 20 dwells coming from each single sweep. The dwells used for this algorithm are listed in Tables 2.4 and 2.5. Many of the rain sweeps were taken from near-tornadic or tornadic locations, making differences in spectral characteristics more likely to be related to scatterer properties. Since this algorithm requires the computation of a vast number of Fourier transforms, a parallelized version of Bootstrap DPSD was developed for this step.

## Chapter 3

### Results

#### 3.1 Spectral Component Convergence Results

We can evaluate the quality and sensitivity of spectra to input parameters using the methods described in Section 2.3.3. By evaluating separately a varied set of spectra for both debris dwells and rain dwells, we can also assess the quality of spectra for dwells with different levels of spectral variance, as debris dwells tend to have higher spectral variance (Bodine et al. 2014; Schneider et al. 2025).

Spectra in Figures 3.1 and 3.2 show characteristics explored in Chapters 1 and 2. They exhibit discontinuities in range due to being taken from multiple sweeps.

The rain spectra in Figure 3.1 have lower spectrum widths and higher polarimetric values, with  $s\tilde{Z}_{DR}$  primarily between 2 and 6 dB and  $s\tilde{\rho}_{HV}$  very close to 1. Spectral differential reflectivity has somewhat higher spectral variance ( $\sigma_{s\tilde{Z}_{DR}}^2$ ) as it appears “speckled”. On the other hand, while  $s\tilde{\rho}_{HV}$  has “speckling” in some dwells, it also has dwells which have segments very low spectral variance ( $\sigma_{s\tilde{\rho}_{HV}}^2$ ), represented by solid red.

Conversely, the debris spectra in Figure 3.2 have higher spectrum widths. For polarimetric variables,  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$  are both lowered and are significantly “speckled,” which is expected, as lower polarimetric spectral components with more spectral variance are expected for debris.

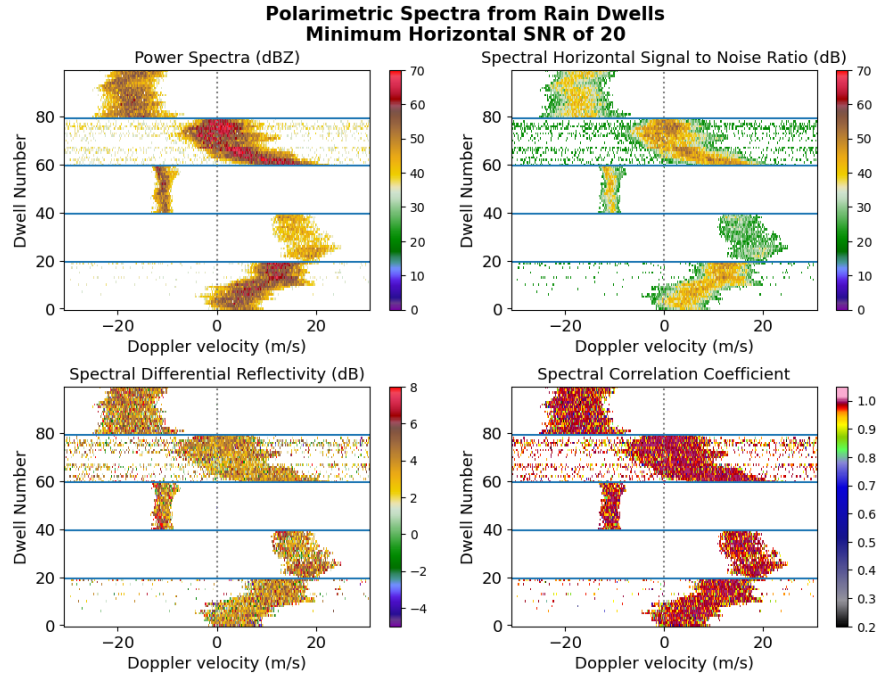


Figure 3.1: Range-Doppler plots for the concatenated rain dwells (separated by blue lines) listed in Table 2.5. The top row (00:19 UTC) in Table 2.5 corresponds to dwells 0 through 19 in the range-Doppler plots, and so on.

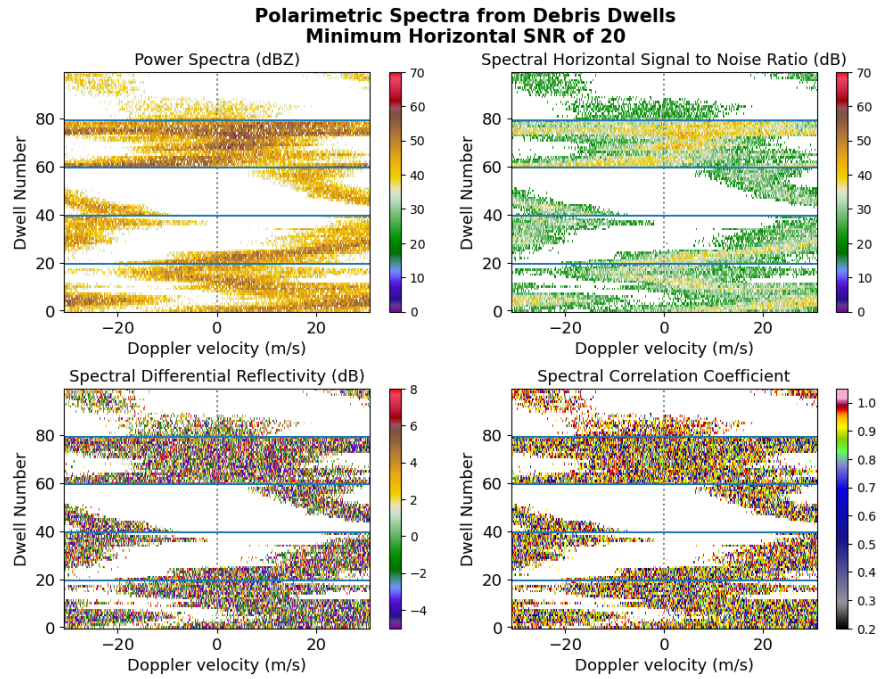


Figure 3.2: As in Figure 3.1 but for the concatenated debris dwells in Table 2.4.

With these spectra, for debris and rain, we run the convergence algorithm to optimize for two parameters, the number of bootstrap samples and the number of pulses in a dwell, which is approximately analogous to the antenna rotation rate.

Figure 3.3 shows the results of the convergence algorithm. Figures 3.3 A, B, and D show similar patterns. Spectral correlation coefficient for rain (Figure 3.3C) has a much greater percentage of unconverged spectral components. For all variables and types, at any given constant dwell length, convergence improves with diminishing returns as the number of bootstrapped pseudorealizations increases.

There are a few weaknesses of the algorithm. One of these is in an assumption about DPSDs with short input timeseries. For rain,  $s\tilde{\rho}_{HV}$  has a lot of unconverged spectral components, with more than 20% unconverged spectral components, even with the highest dwell length and most pseudorealizations. This is likely because the assumption that there will be large changes in values for small numbers of pseudorealizations and small dwells is limited to fields with higher spectral variance. If spectral components of  $s\tilde{\rho}_{HV}$  do not vary significantly with a low bootstrap count or dwell length,  $\max(\Delta s_{k,N[3,\dots,10]})$  may be too low for subsequent iterations of  $\Delta s_{k,N[11,\dots]}$  to become less than 10% of  $\max(\Delta s_{k,N[3,\dots,10]})$ . As seen in Figure 3.4A, many values of  $\max(\Delta s_{k,N[3,\dots,10]})$  are quite low, below 0.005. This is corroborated with the areas of unconverged spectral components in Figure 3.4A. Some of these regions also correspond to areas of low  $\sigma_{s\tilde{\rho}_{HV}}^2$ .

Another weakness of the algorithm is that we have assume that the spectral components converge to a ‘true’ value. It is possible that we are measuring convergence to biased spectral components. These concerns are mitigated by the observation that the computed spectra for debris and rain align with expected behavior based on prior simulations, discussed in Section 1.4.3.2. For debris,  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$  have lower values

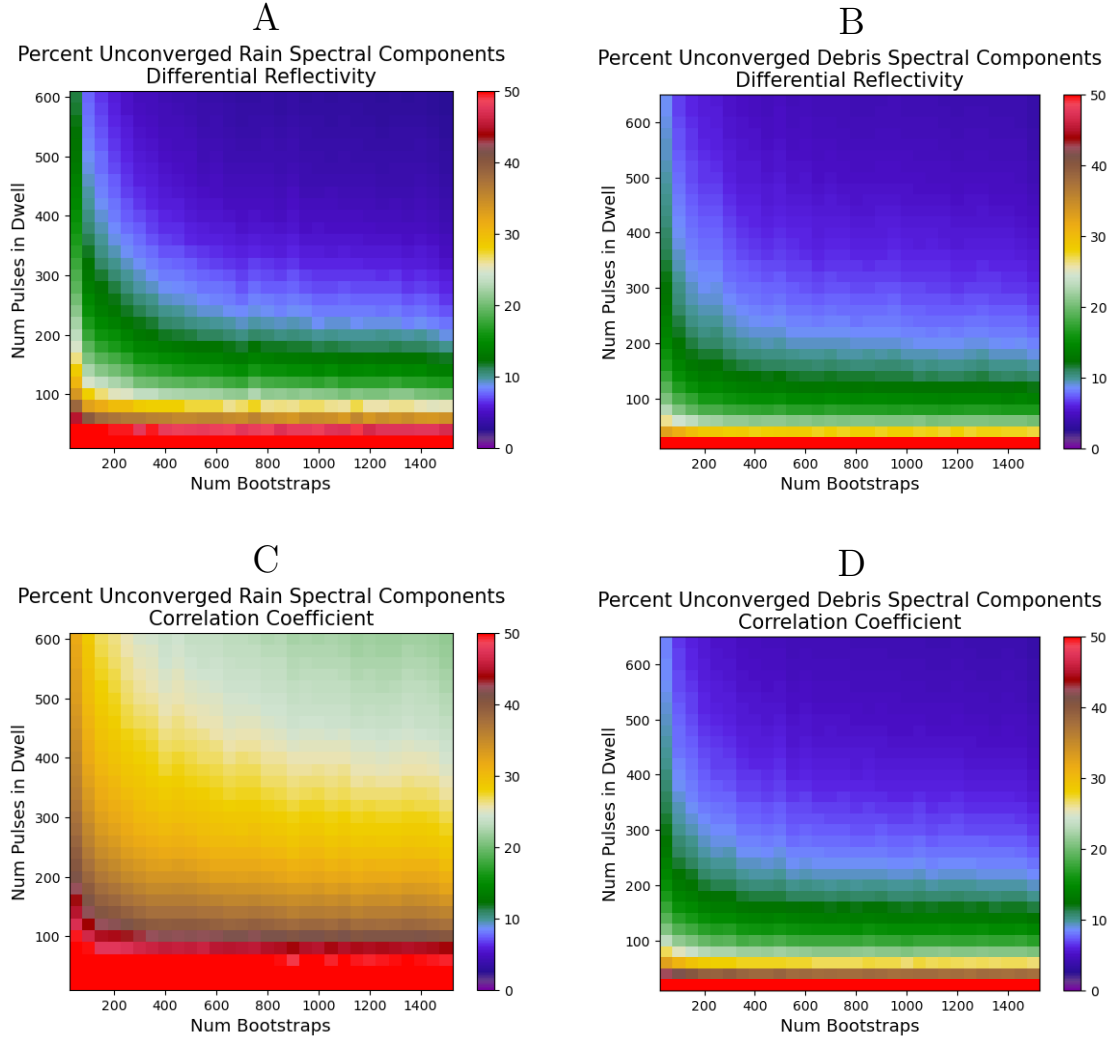


Figure 3.3: Convergence algorithm results for  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$  for both debris and rain. The percent of spectral components for each product remaining unconverged is color coded as a function of both the number of bootstrapped pseudorealizations and the number of pulses used.

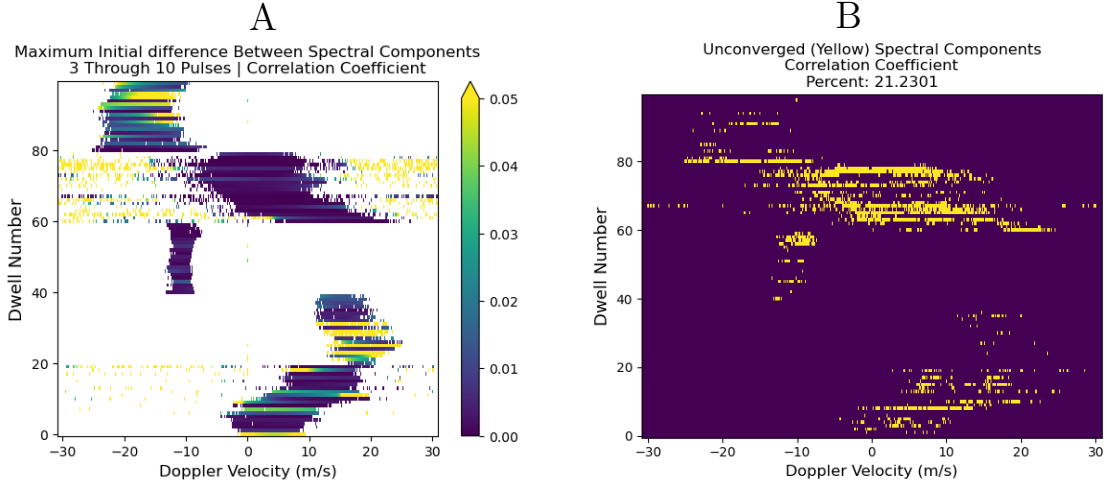


Figure 3.4: A) Values of  $\max(\Delta s_{k,N[3,\dots,10]})$  for  $s\tilde{\rho}_{HV}$  in rain dwells. B) Locations of “unconverged” spectral components for  $s\tilde{\rho}_{HV}$  in rain dwells.

and higher spectral variance, as expected from simulations and consistent with bulk variables.

Finally, many of the values chosen for spectral convergence are arbitrary. For example, the threshold of 10% improvement and 3 consecutive iterations specifically of  $\max(\Delta s_{k,N[3,\dots,10]})$  below that threshold can be tuned such that the percentage of converged spectral components changes significantly. However, it is clear from the Figure 3.3 that for a constant number of pseudorealizations, there are diminishing returns for increasing the length of a timeseries. For rain spectral differential reflectivity and both debris DPSDs, A reasonable point of diminishing returns is around 10% unconverged pixels.

In order to achieve lower than 10% of spectral components unconverged, the length of the dwell should also be greater than around 300-400 pulses. At this number of pulses, greater than 500 bootstraps is also necessary to reach lower than 10% spectral components unconverged. It is also worth noting that with smaller dwell lengths, the number of necessary bootstraps appears to be lower, due to the limited number of

unique bootstraps. At higher bootstrap counts, pseudorealizations are counted multiple times, resulting in diminishing returns.

Despite the limitations above, this test provides important insights. At X-band, to get quality spectra with fewer unconverged spectral components, sacrificing temporal resolution is likely necessary. With around 300 pulses per degree necessary, at a PRT of 0.25 ms, the dish rotation rate would need to be  $12\text{-}15^\circ \text{ s}^{-1}$ . Unfortunately, this does not allow for rapid scanning, and if spectral analysis is desired, the scanning strategy must reflect that.

For operational radars in the NEXRAD program, this problem is magnified. The NEXRAD radars, with a VCP used during rapidly evolving severe weather events, can transmit around 60 pulses per degree (ROC 2025). For RaXPOL, as shown in Figure 3.3, even with increased bootstrapped pseudorealizations, this would yield a high proportion of unconverged spectral components. Furthermore, for a Fourier transform with an equivalent frequency resolution, the Doppler resolution, and thus the ability to distinguish scatterers at similar velocities, changes. Since NEXRAD radars are S-band, their carrier frequency is lower than RaXPOL's X-band frequency. Although this frequency provides a larger Nyquist velocity, reducing the impact of aliasing, the Doppler resolution in a Fourier transform would worsen. Thus, the scanning strategy is again a core limitation of this type of analysis, and even standard-speed scanning is not sufficient for detailed spectral analysis from NEXRAD radars.

Another potentially significant outcome is the tendency for debris spectral components to converge with increasing length of time series, regardless of spectral variance in the frequency domain. Although debris  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$  appear to be quite noisy and speckled in Figure 3.2, the configuration of peaks converges provided that we have enough bootstrapped pseudorealizations, suggesting that this variance is not noise. This is shown in Figure 3.5, where spectra are computed on a full-length debris signal



(the first dwell in Figure 3.2) of 645 pulses and a 450-pulse subset. Although both curves vary strongly across the frequency domain, they vary together with a high correlation, as shown in Figure 3.6. If this spectral variance for debris were induced by signal processing methods, there would be limited co-location of peaks in the spectra as more data were added to the timeseries. This suggests that the cause of spectral variance may be physical rather than associated with signal processing. This spectral variance likely not caused by chaotic motions near the tornado, as rain dwells near the tornado (some of which had high enough winds to be aliased) did not have high spectral variance like debris.

Rather, Bodine et al. (2014) discuss the concept of bulk spatial variance of polarimetric variables in debris. The same arguments from this work may apply, as within a resolution volume, debris has far greater degrees of freedom, i.e., possible sizes, shapes, orientation, angular momentum, and motion. This greater variance in debris characteristics compared to rain could increase both spectrum width and spectral variance. Convergence to specific polarimetric spectra with high variance above around 300 pulses might suggest that this is the Doppler resolution necessary to resolve the scales of these processes.

With these insights, spectra from the Eldorado, OK tornado will be analyzed with time series from the full 1-degree azimuthal dwell length and 1000 bootstrapped pseudorealizations. While 1000 pseudorealizations may be superfluous, it poses no concerns given abundant computational resources and non-real-time processing.

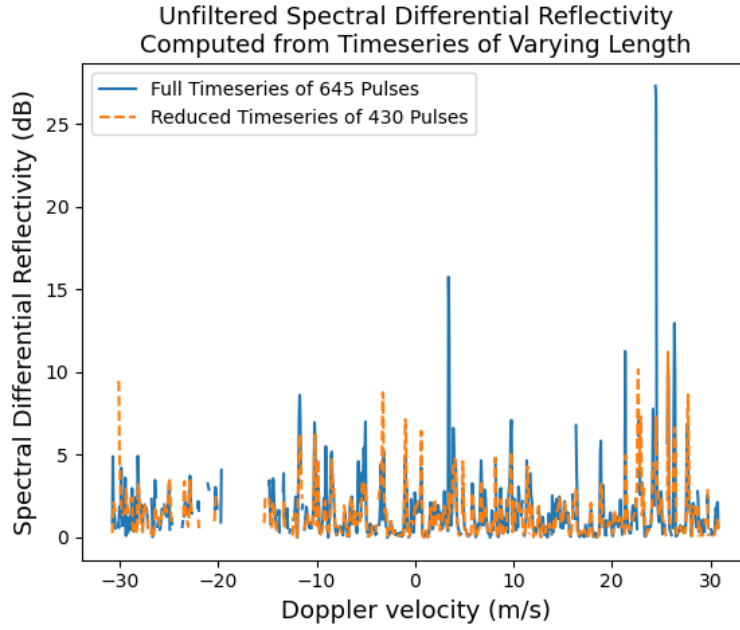


Figure 3.5: Line plots of  $s\tilde{Z}_{DR}$  unfiltered in the frequency domain for two different but still reasonable dwell lengths as determined by the convergence algorithm. The number of bootstrapped pseudorealizations is 1500.

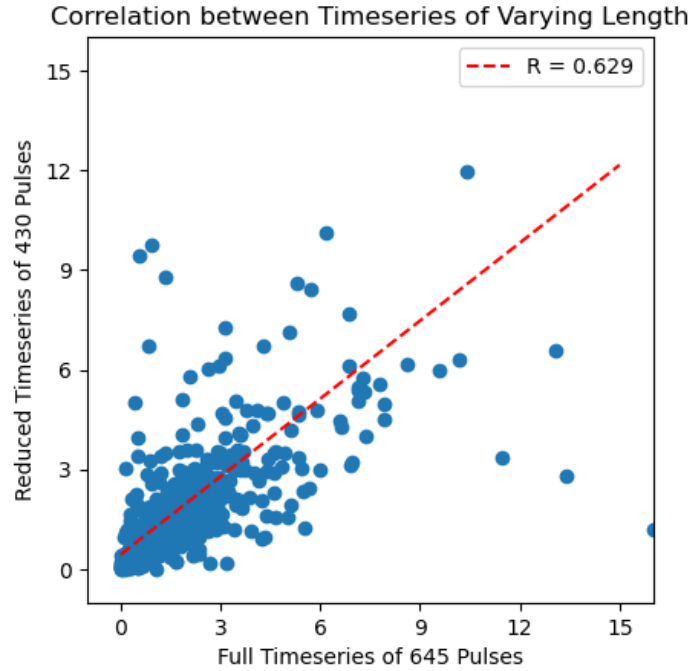


Figure 3.6: Correlation between  $s\tilde{Z}_{DR}$  with full input time series of 645 pulses and subsetting input time series of 430 pulses from Figure 3.5. The Pearson correlation coefficient of 0.629 indicates significant positive covariance between the two series.

## 3.2 Eldorado, OK Spectral Analysis

### 3.2.1 Tornadogenesis

At the time of initiation of the storm near Childress, TX, around 3:00 PM CDT (NWS Norman 2024b), the Childress municipal airport weather station was reporting a surface temperature of 90°F and a dew point of 70°F (NWS 2024). This temperature–dew point spread was marginally supportive of tornadic activity, as liquid condensation levels were higher, stimulating enhanced outflow (Brown and Nowotarski 2019). Initially, the supercell had a more outflow-dominant character, with RFD precipitation near the hook region bowing out.

Prior to tornadogenesis, however, the character of the storm rapidly changed into a more classic shape, with a concave forward flank and large area of strong inflow, as seen in Figure 3.7. Shortly after, a bowl-shaped lowering took shape, and tornadogenesis ensued.

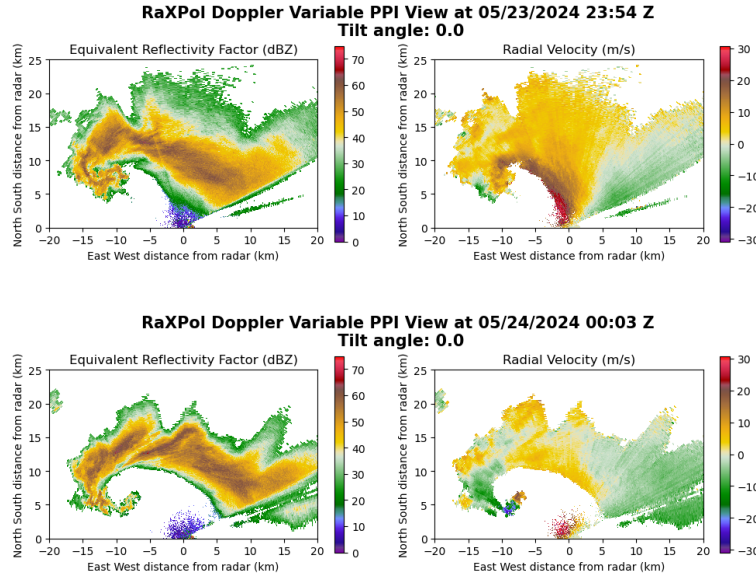


Figure 3.7: RaXPol reflectivity PPI progression between 23:54 and 00:03 UTC, showing the rapid change in character of the supercell’s structure in less than 10 minutes.

Rapid-scan data were collected within the tornado while it remained stationary and intensifying gradually beyond the 10 km range. As the tornado strengthened, it began to drift southeast towards the deployment location. When it came within 10 km of our deployment site, the scanning strategy was switched to slow-scan, single frequency, I/Q data collection.

### 3.2.2 Early Phase Spectral Analysis

Early in the life of the tornado, the bowl-shaped lowering was host to multiple vortices beneath it. Much of the updraft base was rain-free, allowing for the tornado to show minimal scattering save for a ring of elevated reflectivities immediately surrounding the main cyclone.

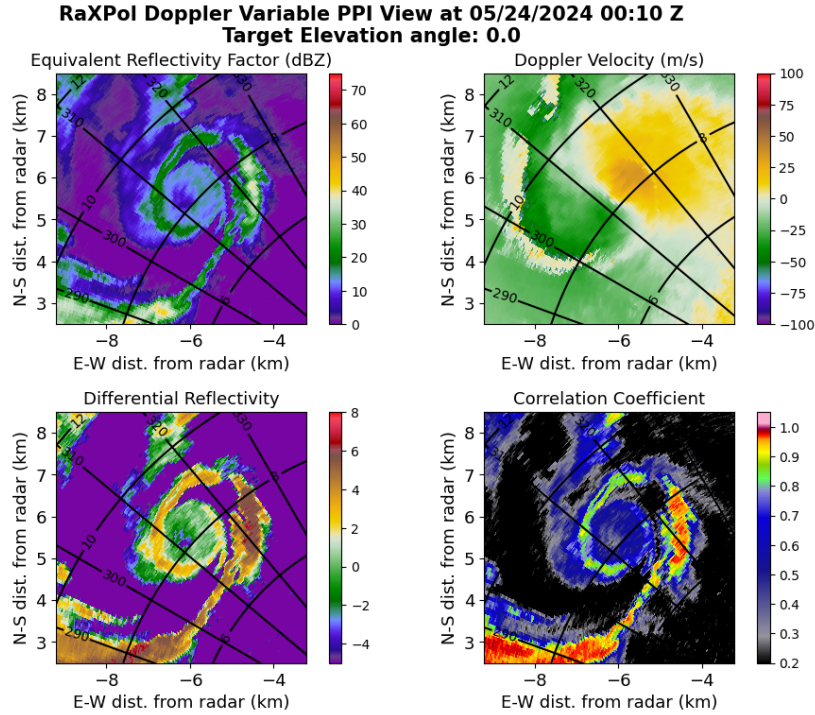


Figure 3.8: RaXPol Doppler variable PPI at 00:10 UTC. Areas of erroneous velocity are also present due to very low reflectivity, and thus, low SNR.

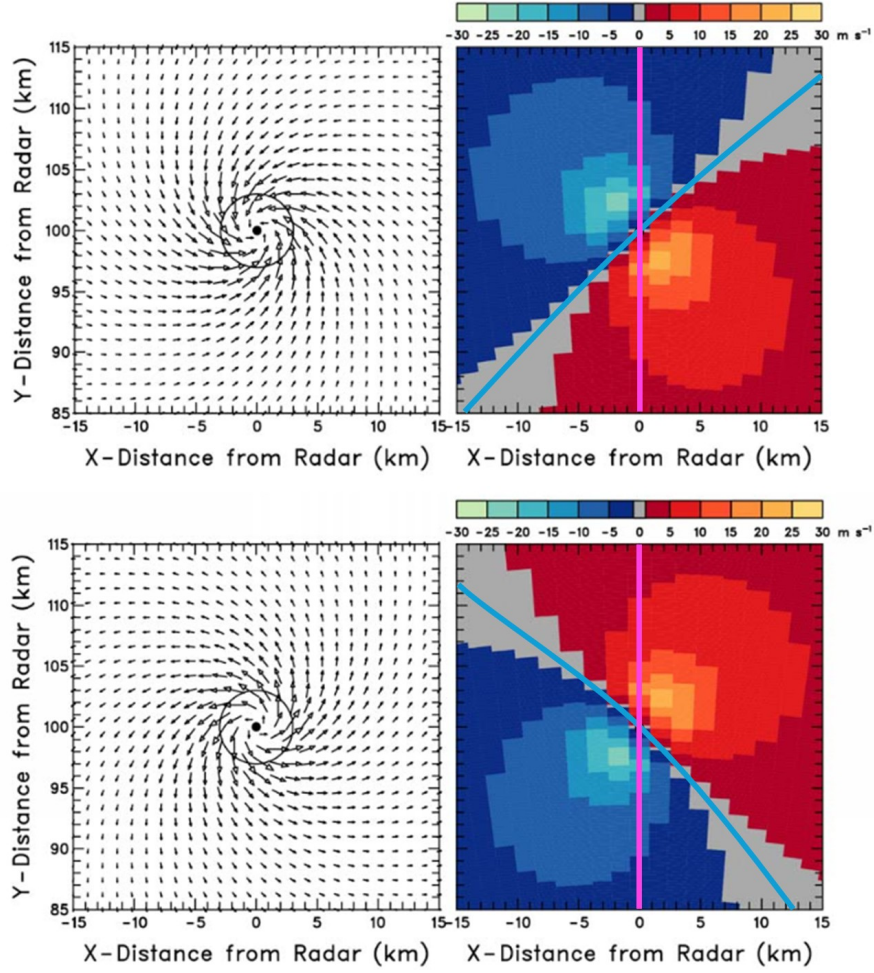


Figure 3.9: Idealized scatterer cyclonic rotational convergent (top) and divergent (bottom) couplets from Figures 4.7.3 (top) and 4.7.4 (bottom) in Brown and Wood (2007). The pink line gives the radial through the center of the circulation and the blue curve follows the zero isodop. In the top plot plots,  $d(az)/dr$  of the blue curve is positive, showing scatterer convergence. In the bottom plots,  $d(az)/dr$  of the blue curve is negative, showing scatterer divergence.

As seen in Figure 3.8, while the presence of a TVS is clear, we cannot ascertain the presence of a TDS regardless of polarimetric characteristics owing to very low reflectivities. Also apparent, are areas near the vortex where there are not sufficient scatterers to provide velocity estimates. The lack of scatterers is especially apparent in the RFD region adjacent to the hook.

In order to diagnose areas of radial scatterer convergence or divergence with respect to the tornado cyclone, as outlined by Brown and Wood (2007), we can observe the change in Doppler velocity with range ( $dv/dr$ ). The top right plot in Figure 3.9 has a negative ( $dv/dr$ ) along the radial highlighted in pink near the core of the cyclone, implying convergence. The opposite is true for the bottom right plot in Figure 3.9.

For cyclonic circulations, we can use a shortcut and compare the slope of the zero isodop (line of constant Doppler velocity) to the radar-centered radial through the center of the tornado vortex. With cyclonic (counter-clockwise in the Northern Hemisphere) rotation, when the slope of the isodop compared to the central radial is positive ( $d(az)/dr$ , where  $az$  is azimuth angle from the radar), this suggests an area of scatterer convergence. The opposite applies to scatterer divergence. When the slopes are equivalent, motion is purely rotational. This method is preferable, as rapid changes in scatterer convergence or divergence will manifest as tight spatial oscillations in the zero isodop, allowing easy qualitative identification. The zero isodop can also be overlaid on other products to correlate with velocity patterns.

This is a common technique in identifying thunderstorm outflow. If the outflow is strong and convergent, wind often reverses direction across an outflow boundary (e.g., southerly to northerly), and the zero isodop often aligns with weak reflectivity echoes. The sign of the slope compared to a radial depends on whether the convergence along the boundary is cyclonic (positive), anticyclonic (negative), or irrotational (zero).

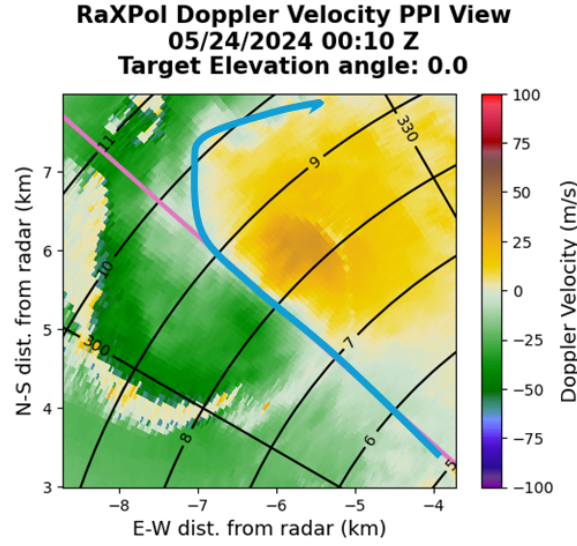


Figure 3.10: RaXPol Doppler velocity PPI at 00:10 UTC. The pink line is the radial through the center of the vortex, while the approximate zero isodop is annotated with the blue curve.

We can apply this technique to the tornado in the early stage in Figure 3.10.

This tornado cyclone at this point was not exhibiting significant scatterer convergence or divergence, signified by the zero isodop parallel to the radial. Outside the tornado cyclone, there is an area of radial scatterer convergence to the north. We expect to see a shallow layer of strong wind convergence immediately outside the tornado’s corner flow, (Lewellen 1993; Dowell et al. 2005; Fischer and Dahl 2022; Rotunno and Bluestein 2024), but at this range, this is not observed.

The measured elevation angle of the sweep was  $0.15^\circ$ . At the range of the tornado, the topography-adjusted beam height above ground level was between 10 and 20 m. With a beam this low, significant ground cluttering should have been observed, as RaXPol’s  $1^\circ$  beamwidth at this range is wider than 15 m. Ground clutter was only observed within 2 km of the radar in the direction of the tornado, as per Figure 3.11. The presence of close-range ground clutter over mostly flat terrain suggests there may have been slight errors in the measured elevation angle. Another factor possibly

compounding to this error is that at sub- $0.5^\circ$  elevation angles, the radar beam can reflect off the ground. At farther ranges, by the time the beam reaches the area of interest, it may be significantly elevated (Snyder et al. 2015). The combination of these errors suggests RaXPol was measuring above the inflow layer at this time.

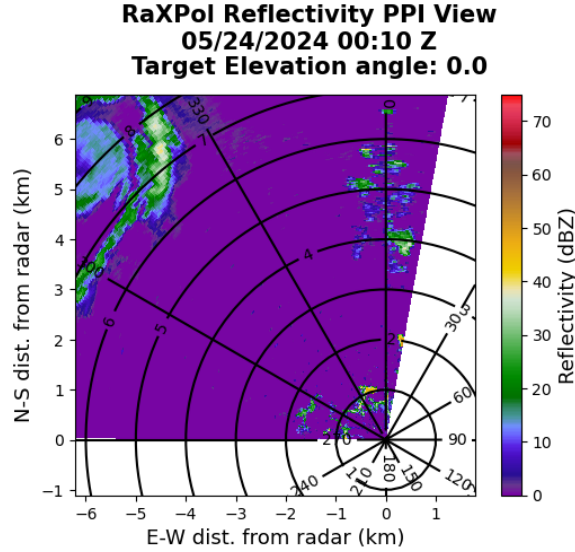


Figure 3.11: RaXPol reflectivity PPI at 00:10 UTC. The echoes in this plot near and north of the radar are ground clutter.

With this in mind, spectra were computed through the center and edge of the tornado. In the spectra in Figures 3.12 and 3.13, the SNRH threshold had to be lowered to 10 dB to view returns consistently. Even so, there were areas, like at around 7.5 km, where returns were too low. It is unclear whether decreased polarimetric variables and increased spectral variance are due to debris or noise. Regardless, the DCA was applied to these spectra.

Figures 3.14 and 3.15 demonstrate the weakness of the DCA at low SNRs. The edges of the spectra, which would normally be masked, are being flagged as debris. Since the velocity correction is a filter applied to the original signal power, these debris



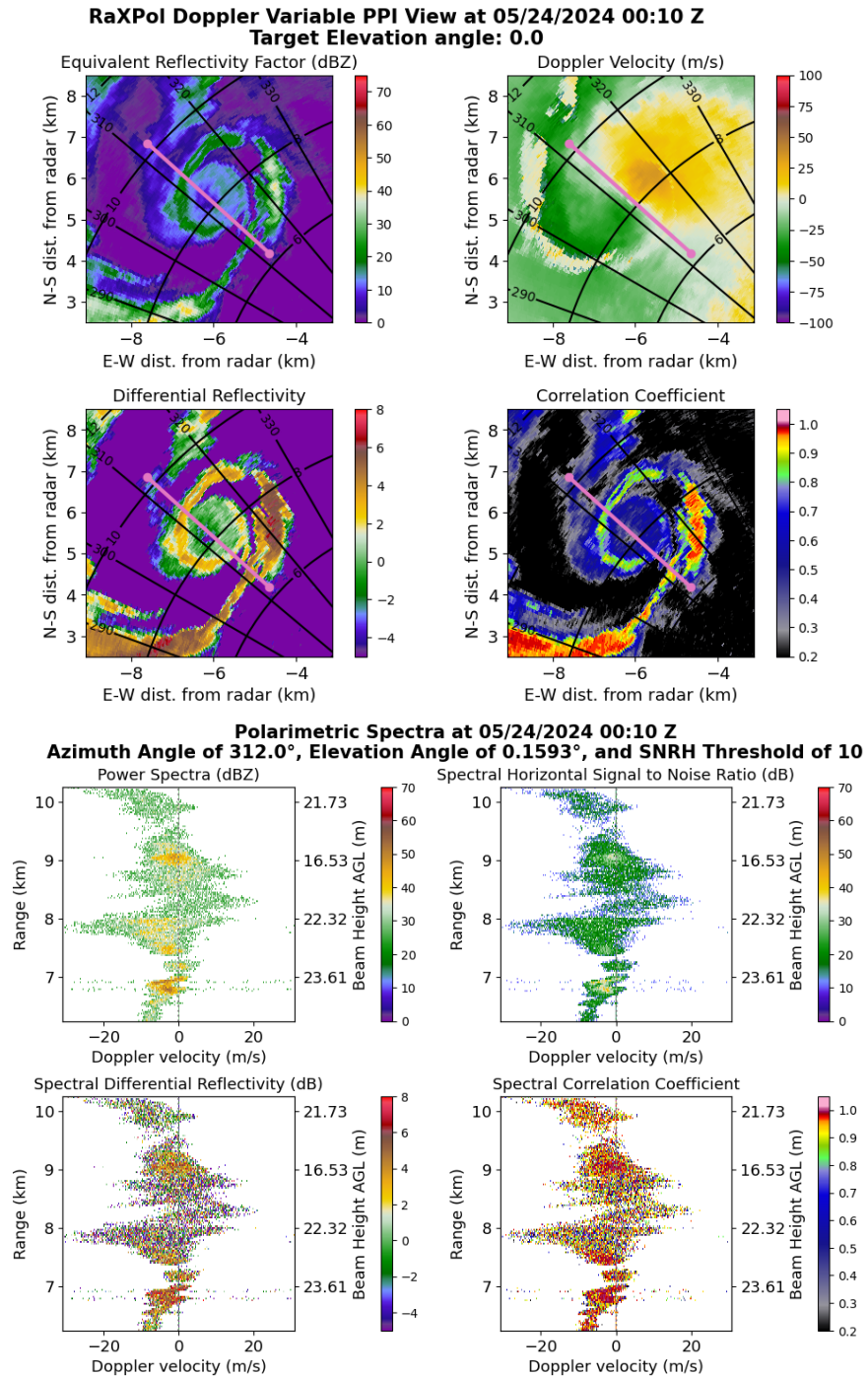
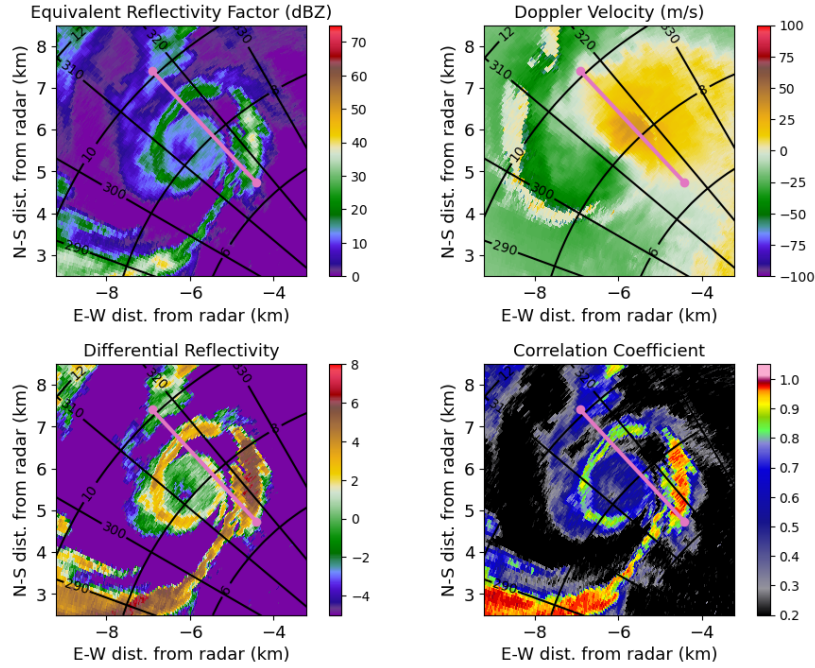


Figure 3.12: RaXPol Doppler variable PPI view at 00:10 UTC (top 4) and accompanied DPSP range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the early stage without significant lofted scatterers.

**RaXPol Doppler Variable PPI View at 05/24/2024 00:10 Z**  
**Target Elevation angle: 0.0**



**Polarimetric Spectra at 05/24/2024 00:10 Z**  
**Azimuth Angle of 317.0°, Elevation Angle of 0.1541°, and SNRH Threshold of 10**

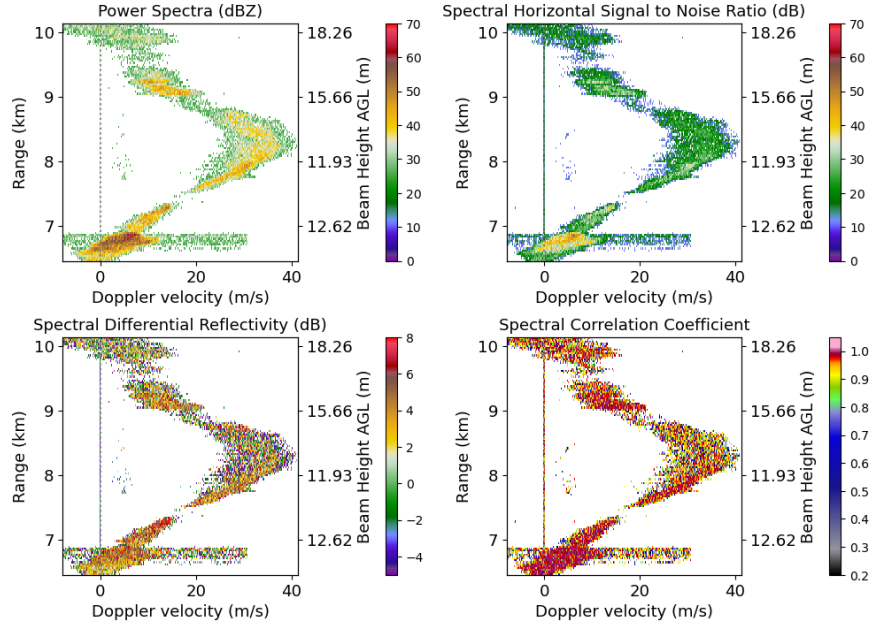


Figure 3.13: RaXPol Doppler variable PPI view at 00:10 UTC (top 4) and accompanied DPDS range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the edge of the early stage without significant lofted scatterers.

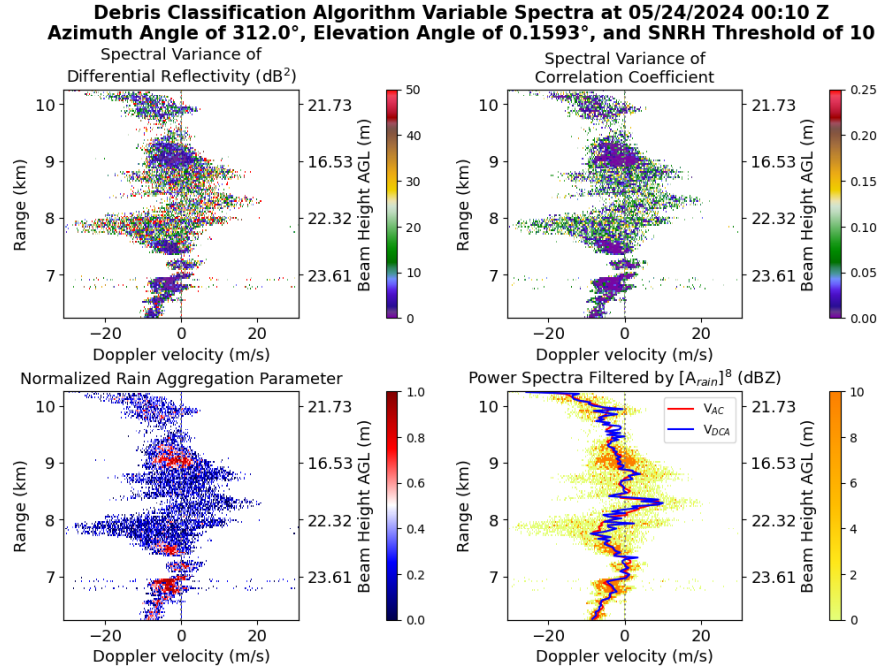


Figure 3.14: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.12 for the Eldorado, OK EF-2 tornado, through the center of the early stage without significant lofted scatterers.

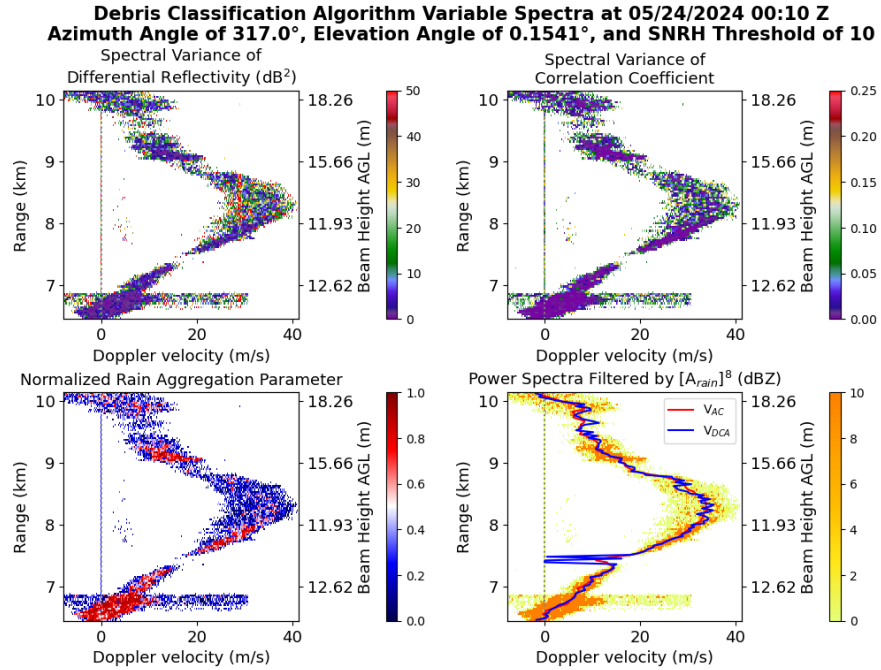


Figure 3.15: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.13 for the Eldorado, OK EF-2 tornado, through the edge of the early stage without significant lofted scatterers.

classifications have a minimal impact on the final corrected velocity. The DCA struggles when only the ground clutter frequency component of the PSDs is above the masked SNR threshold. DCA aggregation filter corrected velocity ( $V_{DCA}$ ) is zero, compared to a non-zero autocovariance velocity ( $V_{AC}$ ), as seen in Figure 3.15.

### 3.2.3 Multi-Vortex Phase Spectral Analysis

#### 3.2.3.1 Tornado Progression During Phase

As the tornado drifted closer to the radar location, it continued to intensify and loft substantial amounts of debris from the ground. This debris was mainly dust and grass, as the tornado had not impacted any structures at this point. As seen in Figure 3.16, within one volume scan, significant debris returns had filled in the WEH. Also, during this intensification phase, the tornado curved back to the north, now poised to miss the deployment location. The tornado's closer proximity to the radar allowed scanning at lower heights above ground level, potentially increasing debris detection.

Additionally, during this phase, multiple vortices were observed on radar, especially on the north side of the tornado. These vortices are identified as concentrated areas of azimuthal shear within the broader circulation and are highlighted in Figure 3.17. Although the temporal resolution of the dataset was not high enough to track evolution of sub-vortices, they were generally observed by the radar to be in the left quadrants with respect to tornado motion, aligning with behavior highlighted in Bluestein et al. (2018).

Between 00:18 UTC and 00:22 UTC, the tornado began a progression toward peak strength. A large cone funnel manifested at the center of the overall broad circulation with vortices surrounding it. Subsequently, the large cone contracted to a rope-like

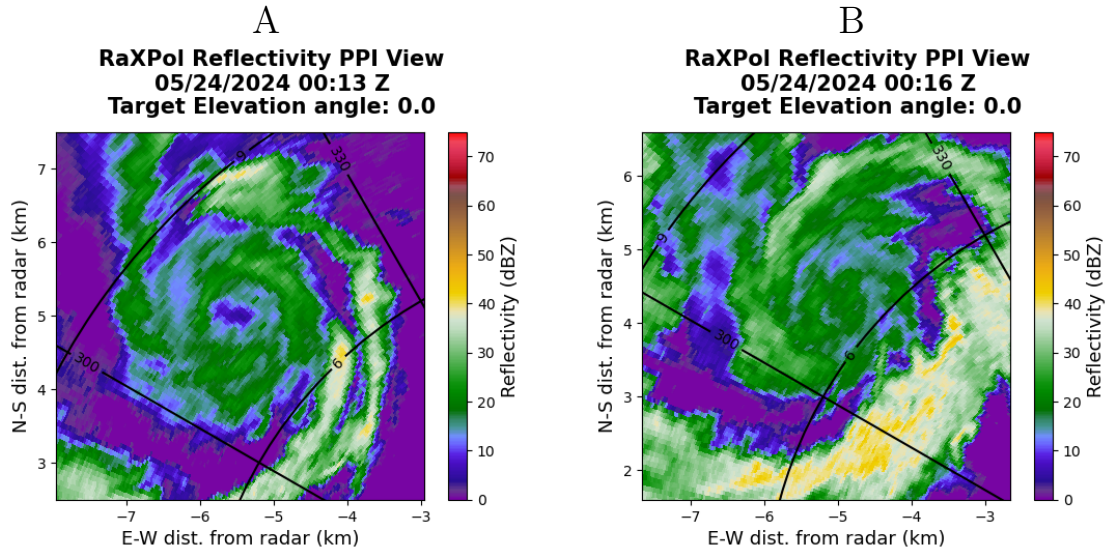


Figure 3.16: RaXPol reflectivity PPIs at 00:13 and 00:16 UTC. Between these two scans, the weak echo hole in the center fills substantially with reflectivity returns due to debris. This suggests the tornado was increasing amounts of debris through this period.

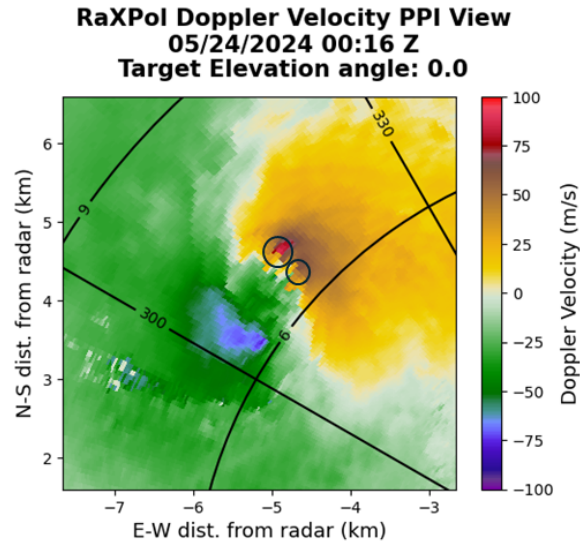


Figure 3.17: RaXPol Doppler velocity PPI at 00:16 UTC. Multiple vortices during this sweep are circled.



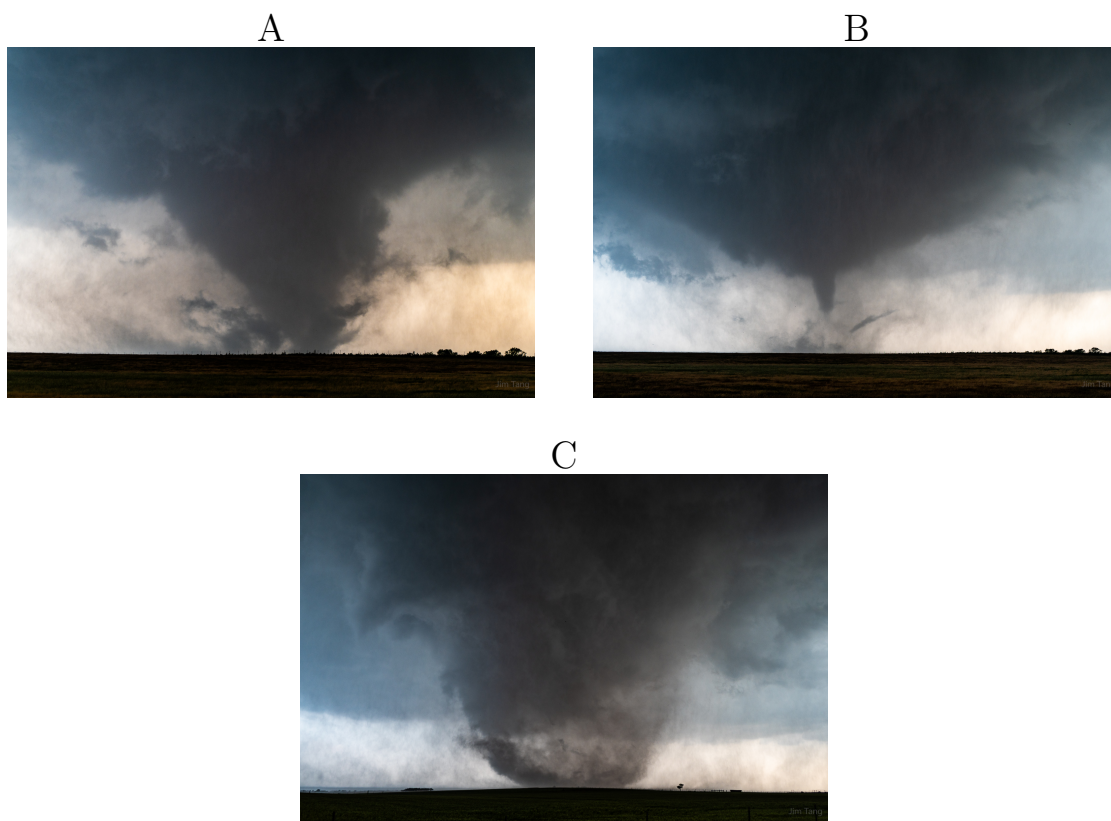


Figure 3.18: The tornado progression photographed by Jim Tang. Photographs were taken from the northeast of the tornado, looking southwest. A) The tornado at 00:19 UTC. A large cone is visible with vortices surrounding it. Debris is observed being centrifuged on the left (southeast) side. B) The tornado at 00:20 UTC. The central rope-like vortex is visible in its later stage with an additional vortex rotating behind it. C) The tornado at 00:21 UTC. The central vortex has disappeared, and the tornado has re-entered a pure multi-vortex stage, albeit with a smaller wind field. The tornado is at peak strength.

vortex in the center of the tornado, as vortices continued to rotate around the tornado at an increasingly smaller radius. Finally, the central vortex broke down into multiple vortices as the tornado reached peak strength. This progression is documented in Figure 3.18.

By the end of this progression, between 00:21 and 00:22 UTC, the tornado reached peak strength, with double aliasing of Doppler velocities. Corrected velocities were in excess of  $95 \text{ m s}^{-1}$ .

### 3.2.3.2 Multiple Vortex Example Spectra

This dataset represents the milestone of the first I/Q data and DPSDs computed through a multiple vortex. In Figure 3.19, the same PPI from Figure 3.17 is highlighted with a radial drawn through a sub-vortex. The DPSDs show an expected result, debris (marked by low  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ ) towards the tornado core and more rain away from the tornado. However, the PSD shows a very small spectrum width outside the multiple vortices (closer than 6.5 km and farther than 6.75 km), suggesting fairly laminar flow. However, in the vortices, spectrum width becomes greater than the Nyquist interval, likely as effective resolution volumes of dwells begin encompassing non-negligible fractions of the sub-vortex structure and potentially as turbulence increases within the vortices.

For these spectra, the spectrum width near the peak wind spans the entire Nyquist interval, and thus, the spectra are impossible to de-alias non-subjectively. In order to de-alias spectra, a discontinuity needs to be created in the power spectra by the SNR threshold. When the whole Nyquist interval is above the SNR threshold, there is no discontinuity present such that it is possible to decide which spectral components need to be de-aliased and which ones can remain.

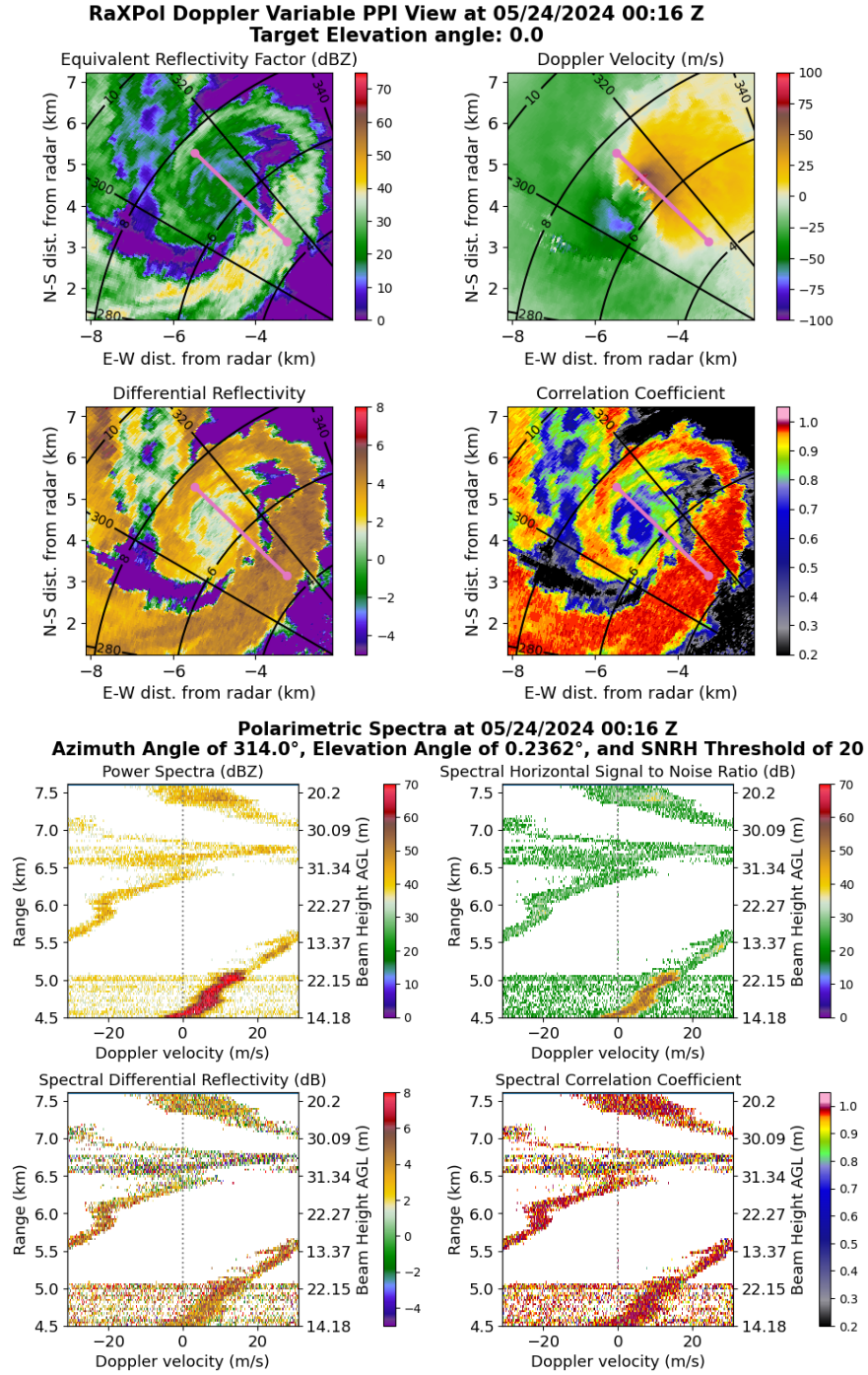


Figure 3.19: RaXPoI Doppler variable PPI view at 00:22 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the peak winds at peak strength at a 2° elevation angle. The blue horizontal line denotes dwell used for single-dwell spectra computation in Figure 3.22.



Next, spectra were computed through the peak winds of the tornado, later in the multiple vortex phase. First, spectra were computed at a higher tilt, as these tended to have lower spectrum widths and could give an idea of the structure we might expect.

In Figure 3.20, the tornado was within 4.5 km of the radar at peak strength. As in Section 3.2.3.1, the reflectivity area was filled such that there was no longer a WEH. Differential reflectivity and correlation coefficient both show depressed values, suggesting that these reflectivities are mostly caused by debris.

The peak winds in this particular scan of the tornado appear to be in the “debris-rain interface” (DRI). The DRI is a region of the tornado marking the edge of debris returns close to the tornado center and start of rain returns farther from the tornado. Data in this area often contains resolution volumes where debris and rain backscatter powers are comparable, and stratification in spectra is most likely here.

In the range-Doppler plots in Figure 3.20 and DCA variable plots in Figure 3.21, stratification is not completely obvious. The DCA also has not significantly corrected velocities for these sweeps. However, if we look at a single dwell at peak winds, as in Figure 3.22, we can see some stratification between debris and rain.

Figure 3.22 displays line plots of the PSD ( $s\hat{S}_H$ ) and the  $An_{rain}$  spectrum in peak wind dwell at the  $2^\circ$  elevation angle in the scan in Figure 3.20. These curves are filtered via the Savitzky–Golay with a window length of 10% of the spectrum width, for human interpretability. Values of the  $An_{rain}$  spectra greater than 0.5 suggest spectral components are more likely to be rain than debris per Schneider et al. (2025).

The highest Doppler spectral components within the main peak of the monomodal right skewed PSD are flagged to be rain per the DCA. Note that this curve has a high mean Doppler velocity, upwards of  $95 \text{ m s}^{-1}$ . This also justifies why velocities were

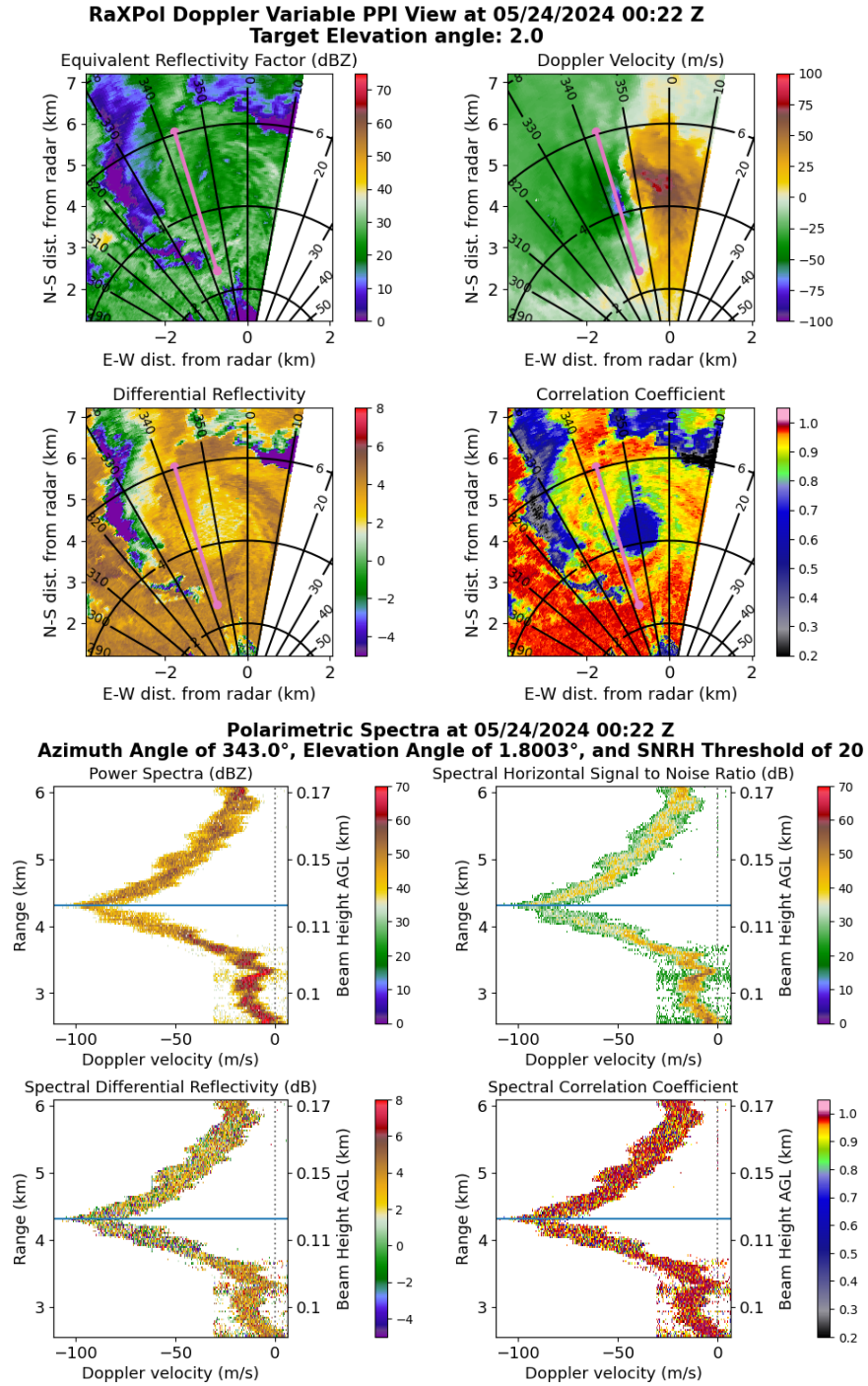


Figure 3.20: RaXPol Doppler variable PPI view at 00:22 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the peak winds at peak strength at a 2° elevation angle. The blue horizontal line denotes dwell used for single-dwell spectra computation in Figure 3.22.

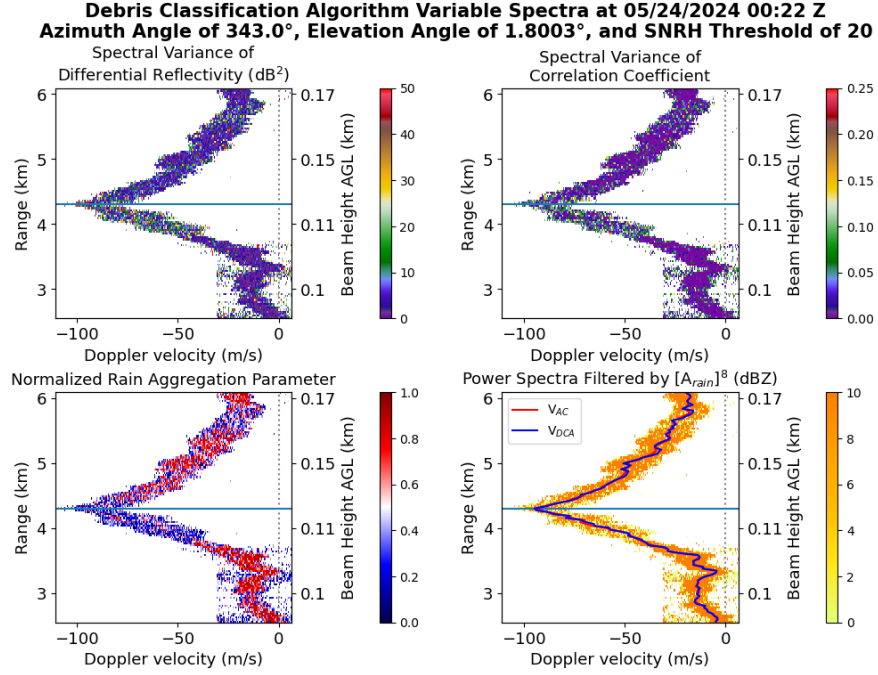


Figure 3.21: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.20 for the Eldorado, OK EF-2 tornado, through the peak winds at peak strength at a 2° elevation angle. The blue horizontal line denotes dwell used for single-dwell spectra computation in Figure 3.22.

not corrected, as the highest PSD components were already classified as rain. These figures show spectra at a higher elevation angle.

At lower elevation angles, below 2°, spectra become ill-defined. In Figure 3.23, the spectra become nearly impossible to de-alias, precluding interpretability. The PPI view plots in Figure 3.23 show a similar picture to the 2° elevation angle in Figure 3.20. The debris cloud of the tornado and peak winds are visible at the DRI. Similar to Figure 3.19, these spectra could also not be de-aliased due to a large spectrum width spanning the Nyquist interval.

This large spectrum width is unlikely to be noise, as there are significant reflectivities, and as stated earlier, spectral components are much above the SNRH threshold of 20 dB. Conversely, it might have legitimate causes, including turbulence in the surface

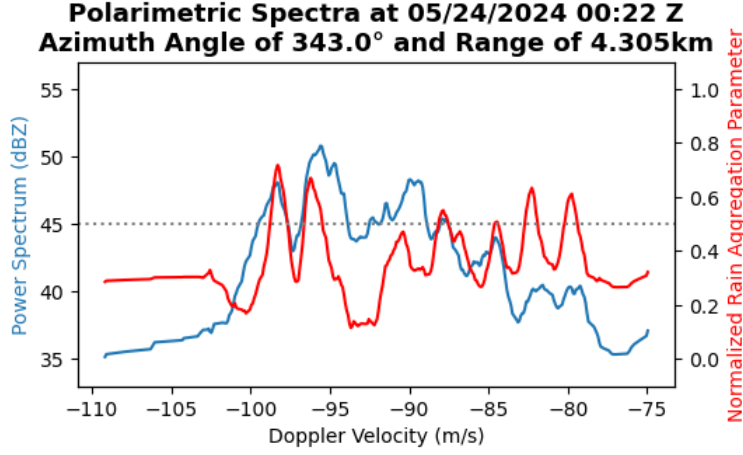


Figure 3.22: Single-dwell spectra for the dwell indicated in Figure 3.20.

layer, strong shearing across the resolution volume, or even differential motion from different debris types/weights. Much of the spectral components have depressed  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ , suggesting this is debris. Regardless of the wide spectrum width, there is still substantial evidence for double aliasing for these velocities close to the ground, as outlined by the black curve approximately overlaid on the spectral peak of the PSDs in Figure 3.23. With increasing range, the spectral peak appears to alias twice before becoming uninterpretable at peak velocities after 4 km, and decreasing back to non-aliased between 4.5 km and 6 km.

As discussed in Section 3.2.2, it is unlikely that the beam height is actually as low as 10 to 15 meters, as there would be a significant ground clutter spike in PSDs and beam blockage over slightly uneven terrain. However, as the tornado was within 4.5 km of the radar at this stage, the velocity field began exhibiting characteristics suggesting surface layer backscatter.

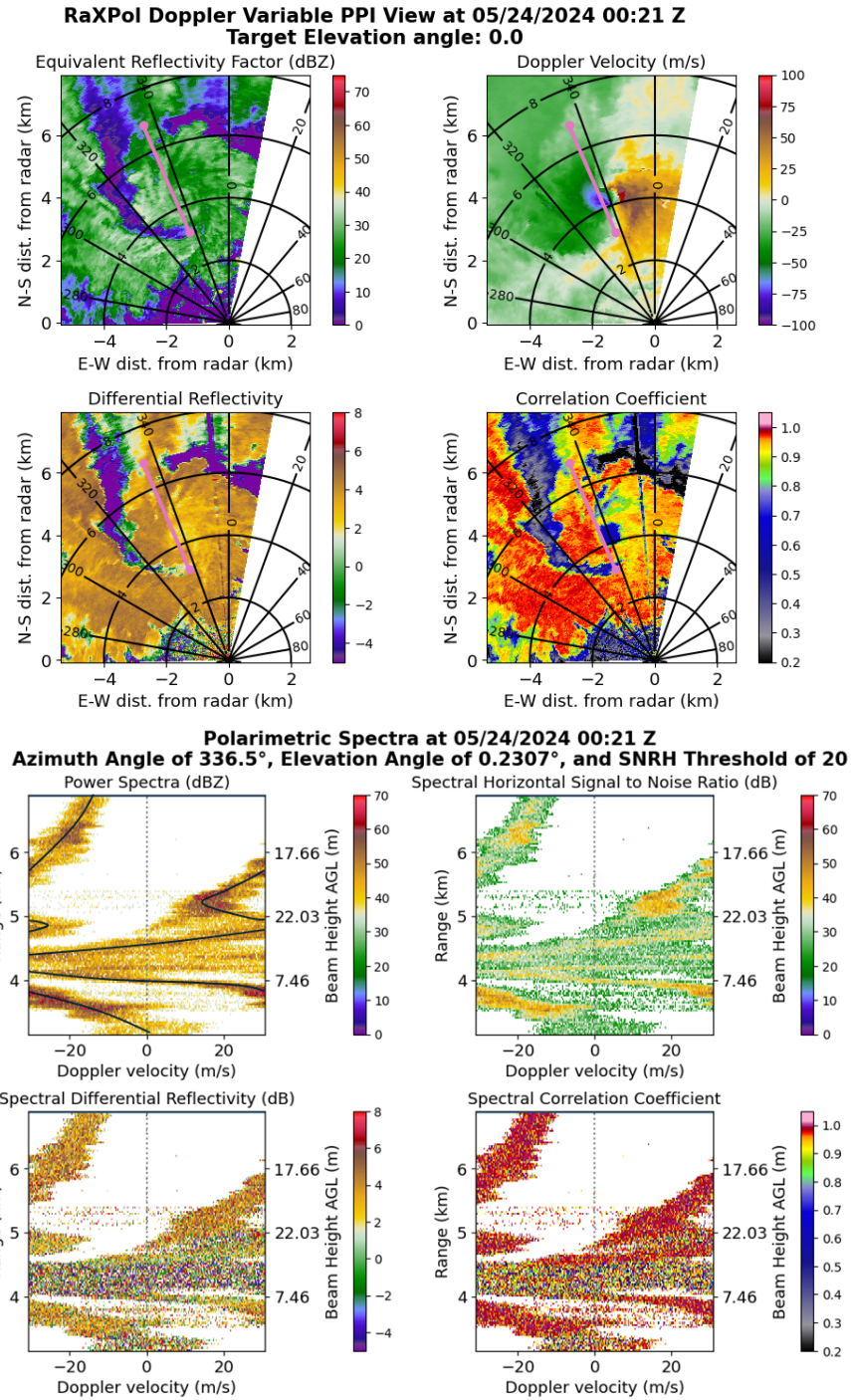


Figure 3.23: RaXPoI Doppler variable PPI view at 00:21 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the peak winds at peak strength at a  $0^\circ$  elevation angle. A black curve is added to the power spectra, representing the spectral peak and showing evidence of double aliasing.

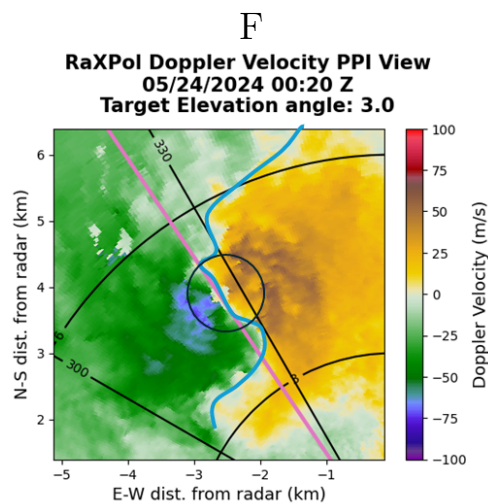
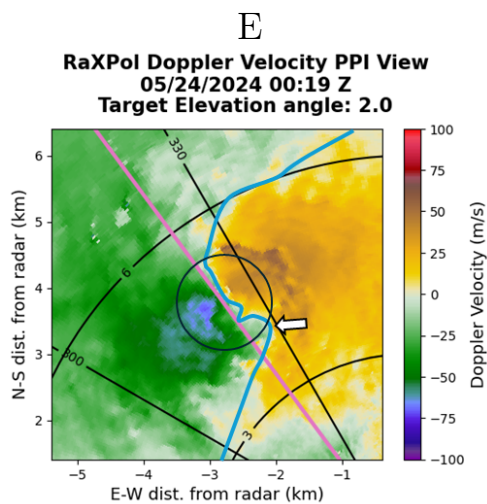
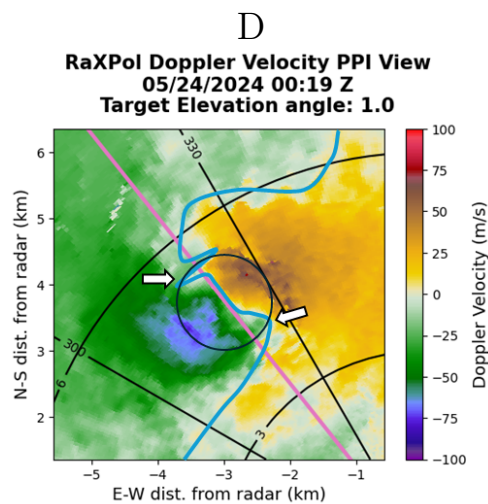
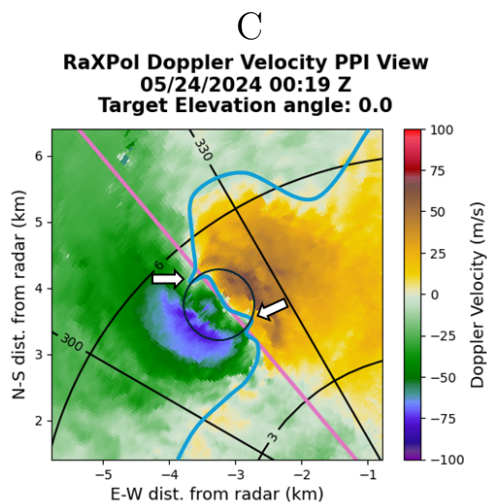
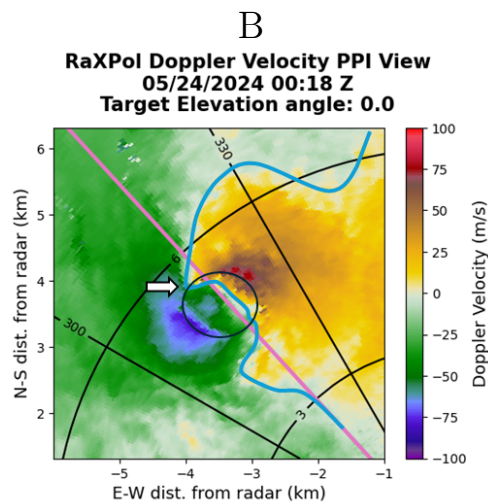
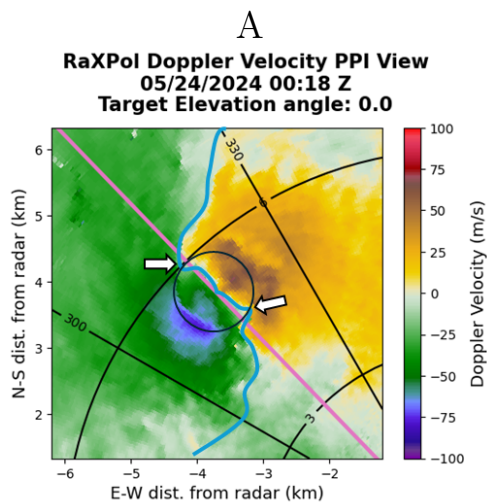
### 3.2.3.3 PPI Analysis

In Figure 3.11, to the north, some weaker ground clutter returns are observed at the same range where the tornado later reached peak strength, between 4 and 6 km. Thus, it is likely that the beam was significantly closer to the ground. The shallow inflow layer discussed in the literature may have been sampled. This is supported by the observation of a wider spectrum width, as observed in the previous section in Figure 3.23, which may be caused by increased shear, increased turbulence, or increased variance in debris sizes, all of which would be more likely to be found near the surface. To determine if this was the case, we can examine the velocity PPI evolution.

Figure 3.24 highlights the progression of the tornado during the period described in Section 3.2.3.1. Initially, the radius of maximum wind (circled) starts broad in Figures 3.24 A through C. However, by the time of the next 0° scan in Figure 3.24I, the radius of maximum wind had nearly halved, and the tornado was at peak intensity, producing winds near ground level over  $95 \text{ m s}^{-1}$ .

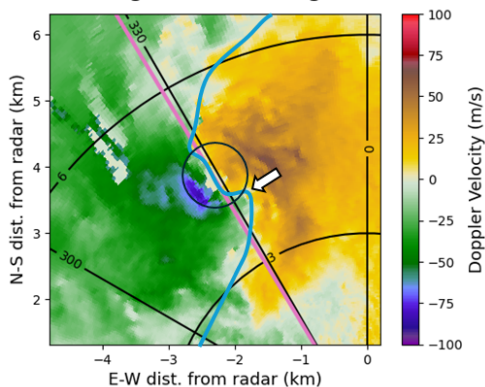
Additionally, regions of strong velocity gradients along a radial can be identified (white arrows in Figure 3.24) near the edge of the tornado vortex, usually at lower tilts. These are represented as sharp discontinuities in Doppler velocity observed along the radial crossing through the tornado center. Typically, strong, radially outward velocities are present inside the tornado boundary, and radially inward velocities are present outside the tornado boundary. This feature is likely caused by strong radial (from the perspective of the tornado) velocity gradients within and in nearby effective resolution volumes, as such, a descriptive name for this signature is the “differential radial velocity couplet” (DRVVC). This feature is almost exclusively co-located with the RMW.





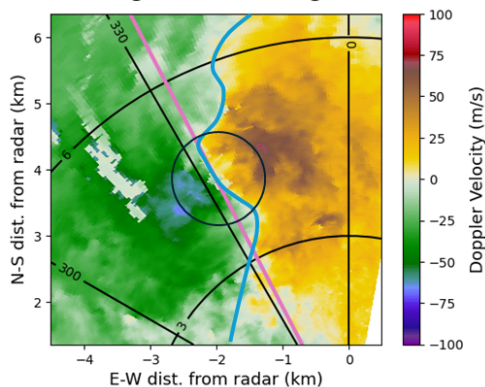
G

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:20 Z**  
**Target Elevation angle: 4.0**



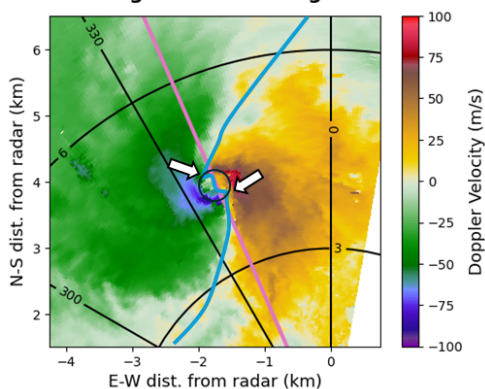
H

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:20 Z**  
**Target Elevation angle: 5.0**



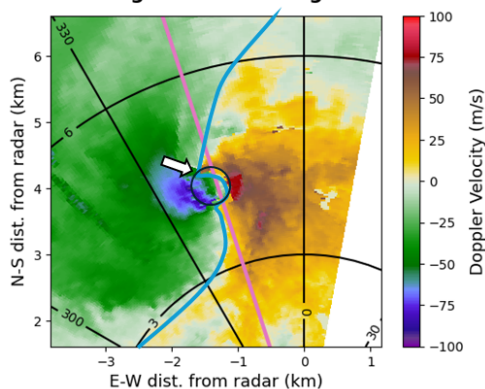
I

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:21 Z**  
**Target Elevation angle: 0.0**



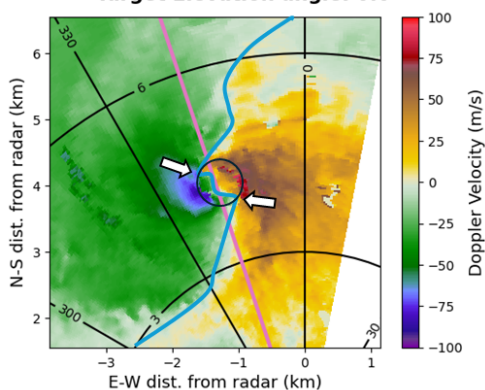
J

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:21 Z**  
**Target Elevation angle: 0.0**



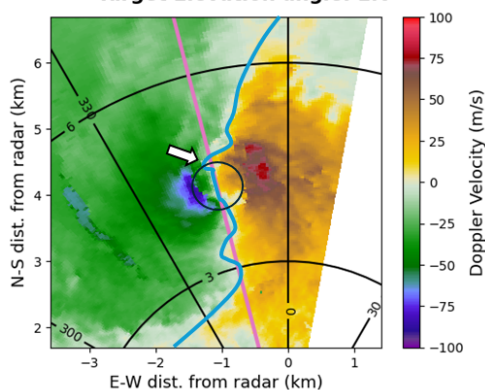
K

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:21 Z**  
**Target Elevation angle: 0.0**



L

**RaXPol Doppler Velocity PPI View**  
**05/24/2024 00:22 Z**  
**Target Elevation angle: 1.0**





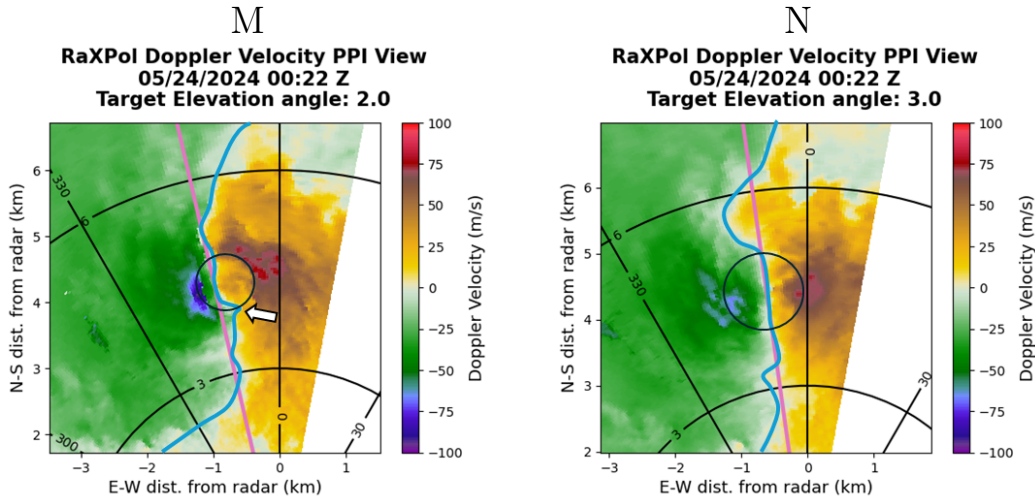


Figure 3.24: Tornado centered velocity PPI progression from 00:18 through 00:22 UTC. As in Figure 3.10, the pink line highlights the central radial, and the blue curve highlights the zero isodop. The black circle highlights the approximate radius of maximum wind (RMW), which is sometimes asymmetric. Additionally, the presence of significant flow discontinuity (and rapid isodop slope changes), is highlighted with white arrows.

Finally, we can compare orientation of the zero isodop to that of the tornado's central radial. Near the surface, at  $0^\circ$  and  $1^\circ$  elevation angles outside the radius of maximum wind, the zero isodop exhibits a positive slope compared to that of the central radial. As discussed in Section 3.2.2, this implies radial scatterer convergence for cyclonic flow. In the DRVC, the zero isodop abruptly changes slope, encountering a "cusp." This slope change is towards scatterer divergence. Within the tornado cyclone, the zero isodop is generally aligned with the central radial, signifying pure rotation. At lower tilt scans, scatterer convergence occurs until it abruptly stops at the DRVC. As such, this represents a data collection success in which the shallow inflow layer modeled by numerical simulations (Lewellen 1993; Dowell et al. 2005; Fischer and Dahl 2022; Rotunno and Bluestein 2024) was actually observed to some degree once the tornado moved within a few kilometers of the radar, as hypothesized by Wakimoto et al. (2012).

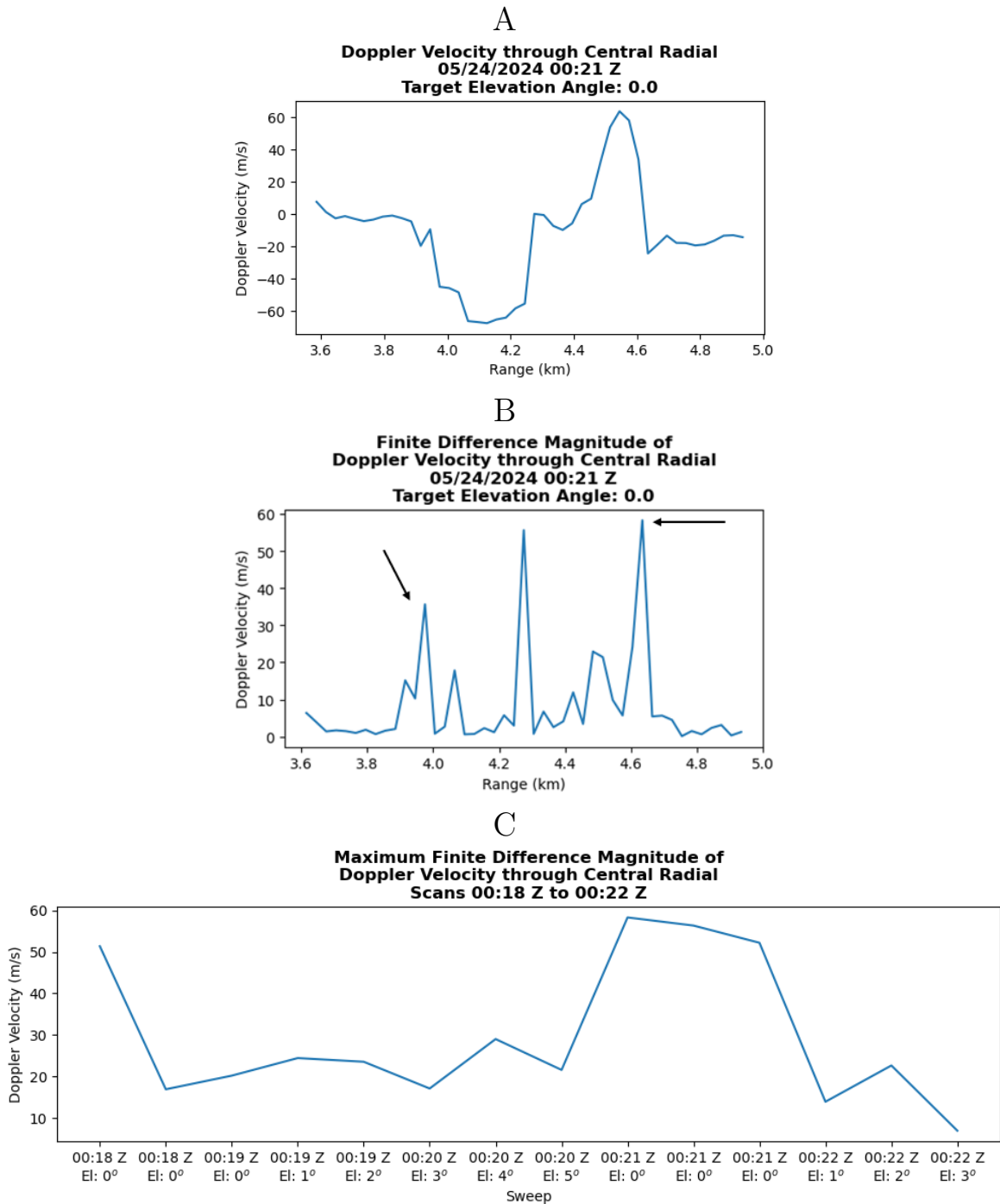


Figure 3.25: A) Doppler velocity through central radial of sweep in Figure 3.24I. B) Finite difference magnitude of Doppler velocity in (A). DRVCs at tornado edge are annotated by arrows. C) Maximum value of (B) from all sweeps in Figure 3.24, showing evolution of strength of convergence/divergence couplets withing tornado vortex.

At higher tilts, DRVCs are usually less pronounced or nonexistent, and scatterer flow is either weakly convergent or completely rotational.

In Figure 3.25, we define an objective measure of DRVCs with height. Since DRVCs are characterized by strong Doppler velocity gradients along a radial, we can compute these gradients (Figure 3.25 A to B). The largest peak at 4.6 km in Figure 3.25B is associated with a DRVC. Plotting the maximum values in adjacent velocity differences Figure 3.25C, we observe that in the first volume, radial scatterer motion gradients do not decrease substantially with increasing tilts. Here, DRVC gradients are similar to those in the tornado core, due to a larger RMW and multiple vortex behavior. Thus, trends are not clearly visible. The pattern of radial velocity gradients weakening with height is apparent strongly in the second volume, starting at 00:21 UTC, after the RMW had decreased substantially. On average, for this whole progression,  $0^\circ$  maximum radial wind gradients were stronger than higher tilts. The observation that the strongest radial wind gradients are associated with DRVCs in  $0^\circ$  scans, supports the claim that DRVCs will tend to weaken with height.

Importantly, the zero isodop cusp associated with the DRVC is co-located with the debris cloud, as seen in Figure 3.26. This implies that not only are tornado dynamics at play when creating this region of strong velocity gradients, but bias from debris centrifuging likely also impacts them. In order to examine the impacts of these velocity biases, spectra were computed in these regions to determine the presence of a debris-induced velocity bias.

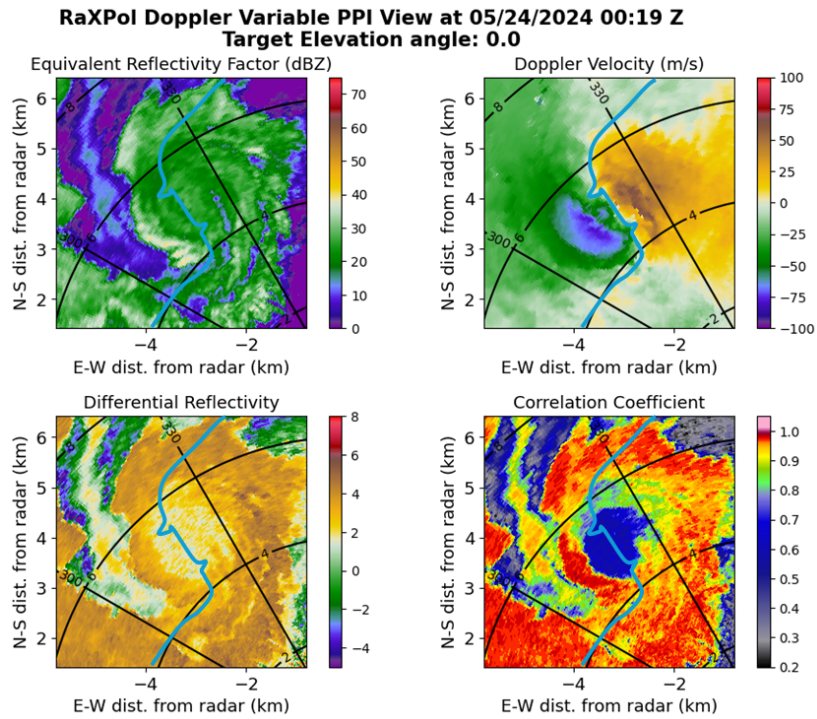


Figure 3.26: RaXPol Doppler variable PPI view at 00:19 UTC. This sweep represents the same time as the sweep in Figure 3.24C. As in Figure 3.24, the zero isodop is annotated by the blue curve.

#### 3.2.3.4 Low-Level Spectra Through Tornado Center

In beginning the spectral analysis of the DRVC region, the first priority is to find possible radial centrifuging. In numerical simulations run with sand, the most similar type of debris to that lofted in this tornado, had the highest difference between the wind and debris motions in the radial component (Bodine 2014). To examine the radial component of the wind in the tornado, the spectra were computed from dwells on the radar radial through the tornado's center.

To increase the chances of finding spectra with both debris and rain in the effective resolution volume of a dwell, a key feature to examine is the area of precipitation wrapping around the west side of the tornado cyclone in Figure 3.26. This area of precipitation is highlighted by higher reflectivities and correlation coefficients. Along the gradients of this precipitation and the debris cloud, dwells were more likely to contain comparable amounts of both. Fortunately, this area of precipitation wrapped around the south of the tornado into a DRVC. Thus, in Figures 3.27 and 3.28, spectra were computed through the center of the vortex along with DCA variables.

In Figure 3.27, the DPSDs clearly denote debris towards the center of the tornado with depressed  $s\tilde{\rho}_{HV}$  and  $s\tilde{Z}_{DR}$ . The DRI closest to the radar is fairly sharp due to the area of precipitation wrapping around. In both DRVCs, the DCA (in Figure 3.28) corrected velocities show lower radial velocities, more towards 0, correcting away from the radial outflow visible in the PPIs in Figure 3.26. This correction strictly occurs in the DRI, where resolution volumes include both rain and debris. Single-dwell spectra were computed to analyze these correction regions.

In both Figures 3.29 and 3.30, there is a clear bimodal distribution in the PSD. The PSD taken closer to the radar shows a more pronounced bimodal distribution.

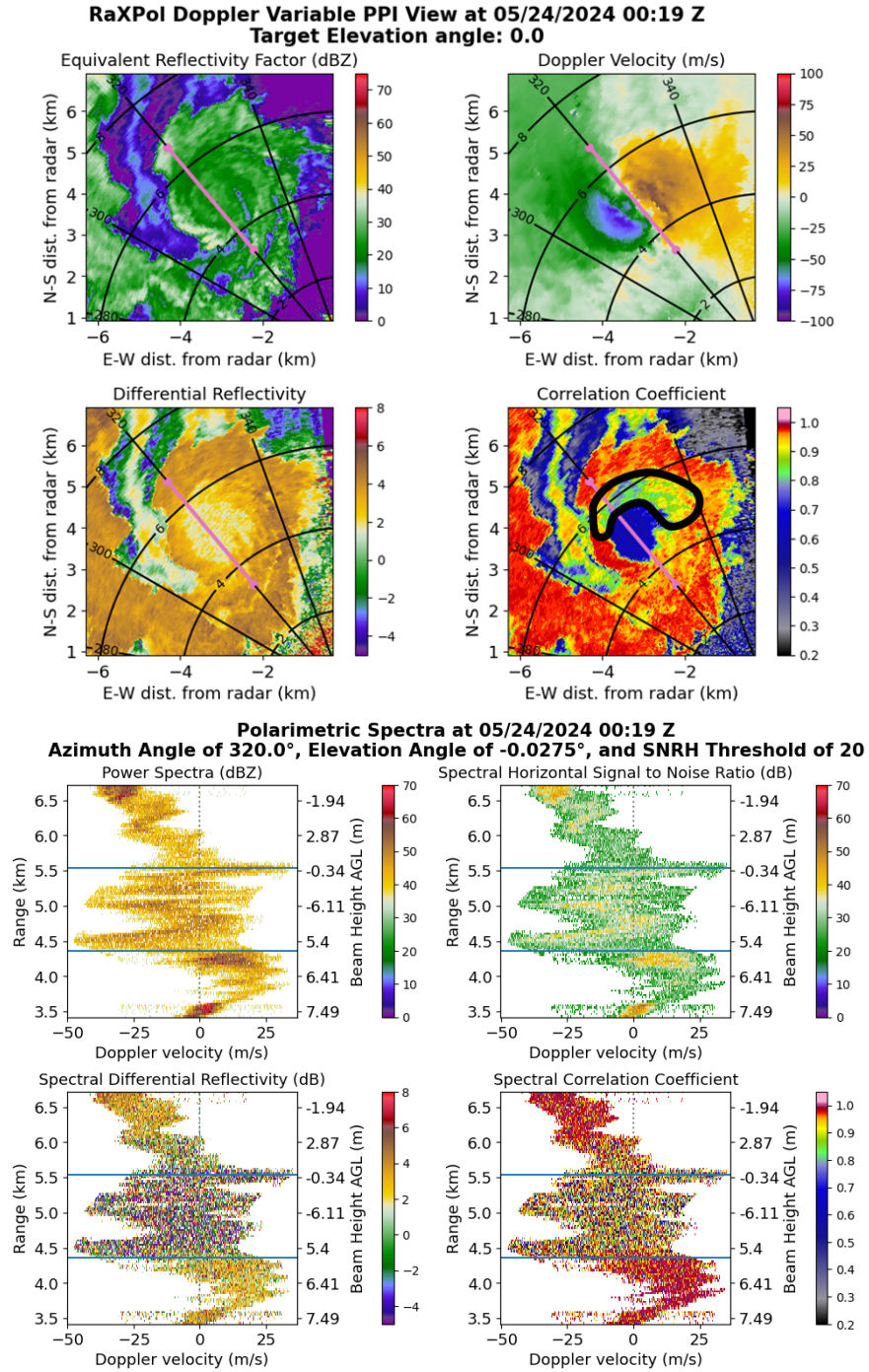


Figure 3.27: RaXPol Doppler variable PPI view at 00:19 UTC (top 4) and accompanied DPSP range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the vortex, traversing the DRVCs. The blue horizontal lines denote dwells used for single-dwell spectra computation in Figures 3.29 and 3.30 and also denote DRVCs of the tornado at this stage. A debris ejection (Kurdzo et al. 2015) is circled in black in the correlation coefficient PPI.

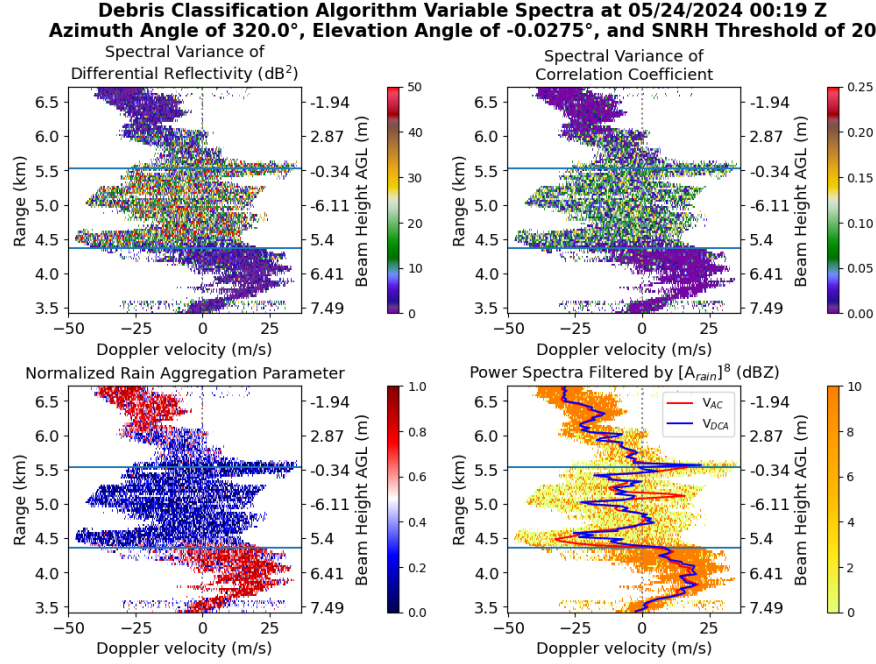


Figure 3.28: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.27 for the Eldorado, OK EF-2 tornado, through the center of the vortex, traversing the DRVCs. The blue horizontal lines denote dwells used for single-dwell spectra computation in Figures 3.29 and 3.30 and also denote DRVCs of the tornado at this stage.

The peak with positive radial velocity has high values of  $An_{rain}$  spectral components, suggesting likely rain, and the peak with negative radial velocity has lower values of  $An_{rain}$ , suggesting debris.

The DRVC in the more distant PSD is much less pronounced. Values of  $An_{rain}$  classify the smaller peak as debris. However,  $An_{rain}$  spectral components in this peak with negative Doppler velocity are consistently slightly higher than the peak with positive Doppler velocity, suggesting that DPSD characteristics lean more towards rain. In Figure 3.27, the correlation coefficient PPI shows the presence of a debris ejection (Kurdzo et al. 2015) rooted near the farther DRVC. Thus, it is possible that this peak may be affected by spectral saturation from debris. Regardless, the DCA still



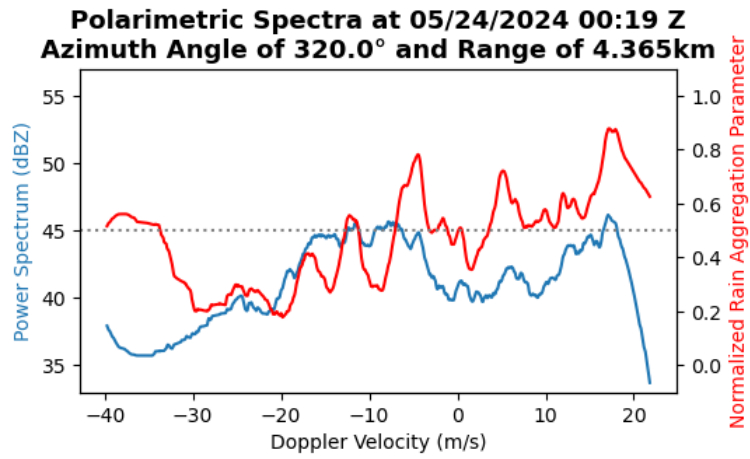


Figure 3.29: Single-dwell spectra for the dwell indicated in the DRVC of Figure 3.27 closer to the radar.

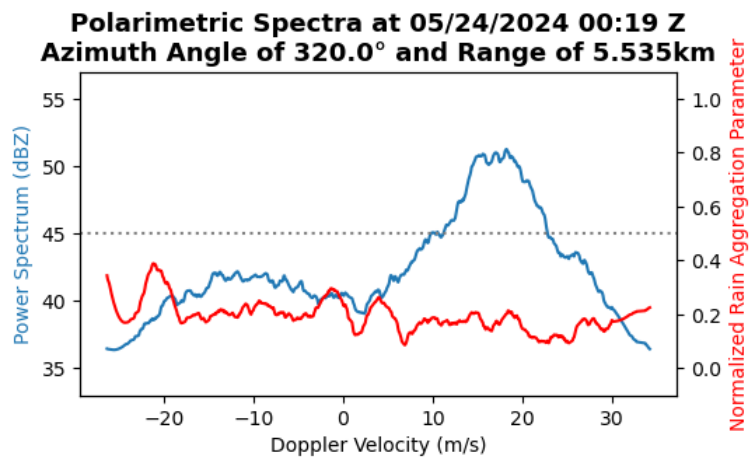


Figure 3.30: Single-dwell spectra for the dwell indicated in the DRVC of Figure 3.27 farther from the radar.



corrected these velocities, as higher values of  $An_{rain}$  likely signify elevated probabilities of rain in those spectral components, even though they were classified as debris.

Unlike the vortex DRVCs, Figure 3.28 demonstrates minimal DCA velocity correction where spectral components are dominated by rain or debris, as limited differential filtering occurs. The limited correction in the debris cloud may be attributed to a combination of two factors. The first factor is spectral saturation. As debris returns become large, their variable polarimetric characteristics can mask rain signals, limiting the efficacy of filtering (Schneider et al. 2025). The second factor likely stems from a limited amount of rain in the core of a tornado, as evidenced by the existence of a weak echo hole aloft. Rainfall is a top-down process, while debris-lofting is a bottom-up process. Near the surface in the vortex core where centrifuging effects are strong, a rain presence comparable to that of debris is not likely as the primary viable source for rain in a vortex is recycling through the shallow surface inflow layer (Dowell et al. 2005).

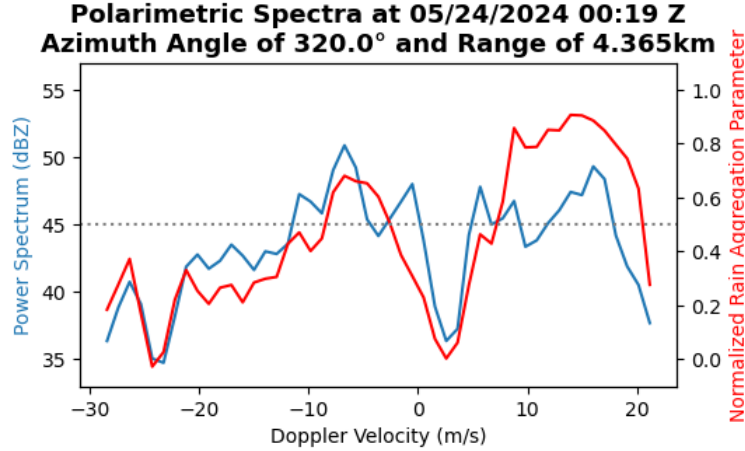


Figure 3.31: As in Figure 3.29, with an I/Q time series of length 60.

Interestingly, as shown in Figure 3.31, the PSD, although very coarse in resolution, still shows a bimodal distribution. With 60 pulses, RaXPOL's antenna rotating at

approximately  $67^\circ \text{ s}^{-1}$  would have produced this spectrum. Although the convergence statistics for time series with this length showed poor convergence as per Section 3.1, for the large and obvious bimodal PSD, the general structure was preserved with the shorter time series length. However, with the decreased frequency resolution, the peak with negative Doppler velocity is misclassified as rain,  $An_{rain}$  greater than 0.5 through most of the peak, in contrast to Figure 3.29 where  $An_{rain}$  is generally lower than 0.5. This may imply that if the DRVC region is being observed in a PPI, the scan speed can be increased while still preserving a bimodal power spectrum due to larger degrees of stratification between debris and rain. However, for more subtle features in the spectra, including the ability resolve of spectral variance in DPSDs to classify spectral components as debris or rain, a slower scan rate is needed.



Figure 3.32: Photograph from Figure 3.18A. The photo is from the northeast looking at the tornado to the southwest. This is approximately orthogonal to the radar beam, which is from the southeast to the northwest. The photograph is annotated to show the approximate location of spectra from Figure 3.29 (magenta circle) and the approximate path of the radar beam (blue arrow). Photographed by Jim Tang.

In addition, Figure 3.32 was taken at the time of the spectra in Figure 3.29 and 3.30. In this photo, the debris cloud is sloped upwards with increasing radius away from the tornado. Since the tornado was moving from right to left from the perspective of this photo, this debris could not have been left behind by the tornado and was ejected forwards. The increased height with radius is also expected, as Dowell et al. (2005) observed, due to the shallow surface layer inflow lifting and deflecting debris upwards. Thus, the circled region in Figure 3.32 contains evidence of debris being centrifuged.

Looking more in-depth at the spectra in Figure 3.29, we are very likely at least sampling some degree of this inflow. The rain peak with positive radial velocity is located at around  $17 \text{ m s}^{-1}$ , and the peak with negative radial velocity is located at around  $-10 \text{ m s}^{-1}$ . This difference between debris and rain motions is quite high, and within the upper end of reasonability per numerical simulations in Bodine (2014), which had maximum radial motion differences between sand and wind on the order of  $20 \text{ m s}^{-1}$ . It is more likely that this large difference in radial motion is a combination of both centrifuging and vertical shear within the effective resolution volume. Within this dwell, the radar is sampling both the shallow inflow layer and the debris centrifuging above it.

To verify additional instances of bimodal PSDs, spectra were computed through the center of the tornado in other sweeps. In Figure 3.33, spectra were computed associated with the next sweep, at  $1^\circ$  elevation (Figure 3.24D).

Figures 3.33 and 3.34 highlight that the DRVCs can occur at higher tilts, too. These velocity discontinuities are co-located with the boundary of the debris cloud. Prior to discussing polarimetric characteristics of the spectra, it is worth noting sharp discontinuity with range present in the PSDs. Specifically, at the 4 km range and 5.5 km range, radial velocity rapidly switches direction from radially inward outside the

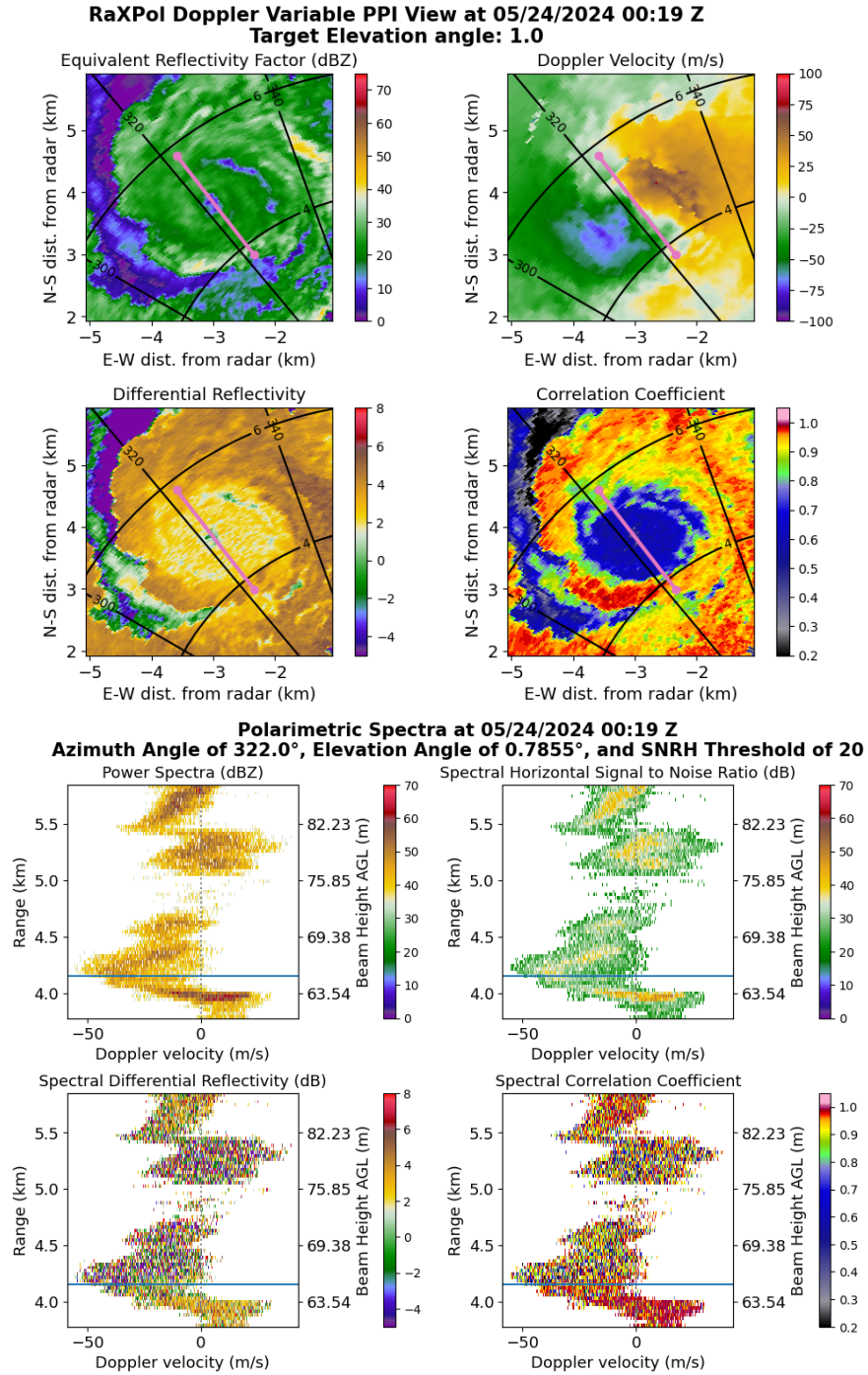


Figure 3.33: RaXPol Doppler variable PPI view at 00:19 UTC (top 4) and accompanied DPSP range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the vortex at a  $1^\circ$  elevation sweep, traversing the DRVCs. The blue horizontal line denotes the dwells used for single-dwell spectra computation in Figure 3.35 and also denotes DRVCs of the tornado at this stage.

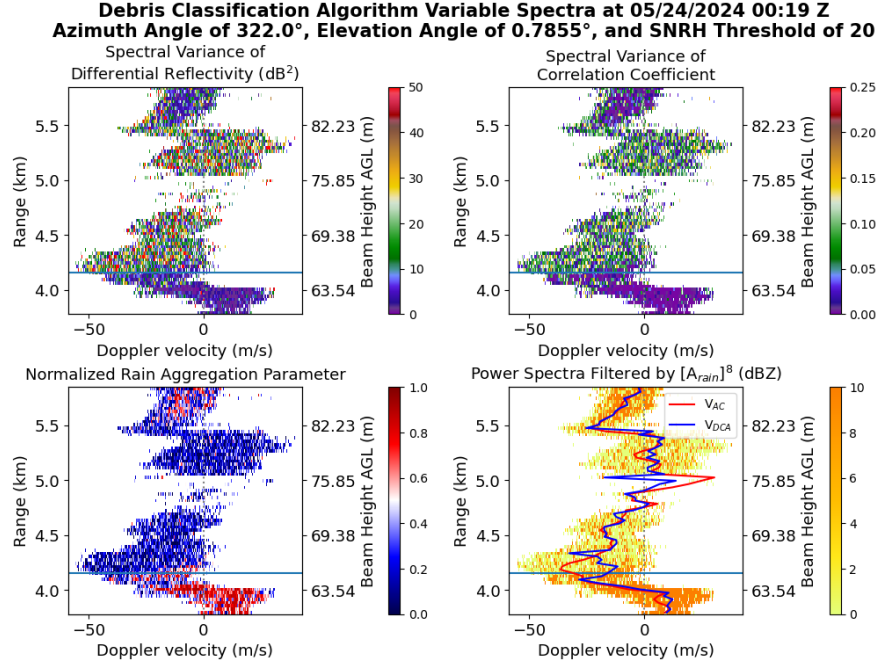


Figure 3.34: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.33 for the Eldorado, OK EF-2 tornado, through the center of the vortex at a  $1^\circ$  elevation sweep, traversing the DRVCs. The blue horizontal line denotes the dwells used for single-dwell spectra computation in Figure 3.35 and also denotes a DRVC of the tornado at this stage.

RMW region and radially outward in the DRVC regions. For this scan, the velocity discontinuity is less than the Nyquist interval, so it is not the result of incorrect de-aliasing. However, in other scans, this discontinuity can approach the Nyquist interval, creating challenging de-aliasing in both spectra and velocity PPIs. In the DRVC closer to the radar, this direction switch occurs mainly in rain signals and is not significantly corrected by the DCA. Deeper in the DRVC in the DRI, however, there are significant velocity corrections of up to  $25 \text{ m s}^{-1}$ , as seen in Figure 3.34. Once again, we can compute single-dwell spectra in these regions.

Figure 3.35 at 4.155 km once again shows a bimodal PSD with the peak corresponding to radially inward velocities for the tornado classified as most likely rain. This time, the difference between the two peaks is lower at  $20 \text{ m s}^{-1}$ , within the realm

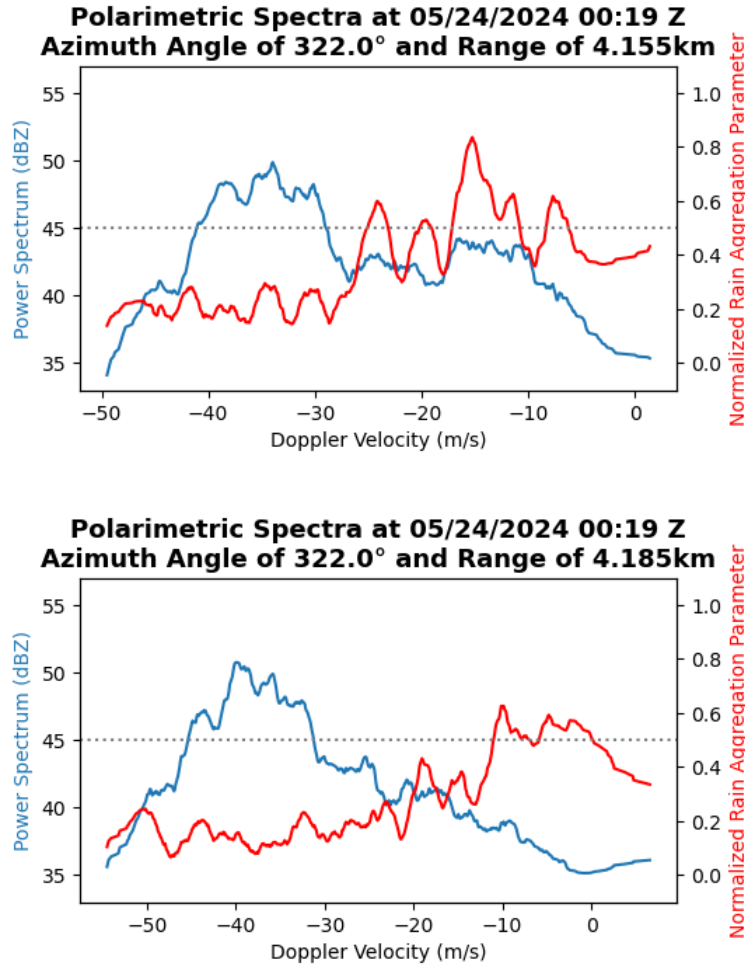


Figure 3.35: Single-dwell spectra for the dwells indicated in Figure 3.33.

of numerical simulations (Bodine 2014), so it is possible that this bimodal PSD could be driven by centrifuging. However, the location in the DRVC means that vertical shearing is likely still ongoing in the resolution volume. It is unknown to what depth this vertical shearing exists due to beam height ambiguities, but these observations are likely evidence of it. Additionally, while some spectra in these DRVCs are bimodal, others are monomodal but very wide, accommodating both debris and rain. More spectra through the tornado at this stage and elevation were computed but not shown for the sake of brevity, and the PSD at the 4.185 km range is an example of this more common type.

Unfortunately, after the radius of maximum wind of the tornado decreased significantly, spectrum widths within the tornado began spanning the full Nyquist range. Thus, analysis of low-level spectra from the subsequent volume was not possible, as the de-aliasing process was far too subjective regarding which spectral components to de-alias and which ones not to de-alias.

### 3.2.3.5 Higher Tilt Spectra Through Tornado Center

Seen in numerical simulations, the maximum magnitude of centrifuging-related errors usually occurs above the corner flow region within the sloped debris cloud, as seen in Figure 3.32 (Dowell et al. 2005; Bodine 2014; Bodine et al. 2016b). Thus, one would expect to observe a lower magnitude of centrifuging at higher elevation angles. Spectra were computed from a higher elevation angle to test this.

At this tilt, we observe a DRVC with a slight velocity discontinuity at the 4 km range in Figure 3.36. At this elevation, the DRI, closer at around 3.65 km, does not appear to be co-located with this DRVC. Single-dwell spectra taken through the DRI in Figure 3.38 highlight the common pattern of a Gaussian PSD with a skewed distribution of  $An_{rain}$ . In this case, again, the lower-magnitude Doppler velocities, thus those with a lower velocity component radially outward from the tornado, are classified as rain. Such spectra are the most common in the DRI regions, where centrifuging effects are apparent but not as strong as in the lower levels. The DCA did not correct these velocities significantly, as the highest rain velocities were already near the spectral peak.

**Potential DRVC Causes** Beginning with Section 3.2.3.3, we introduced the concept of a DRVC and analyzed various spectral properties in the vicinity of these areas.



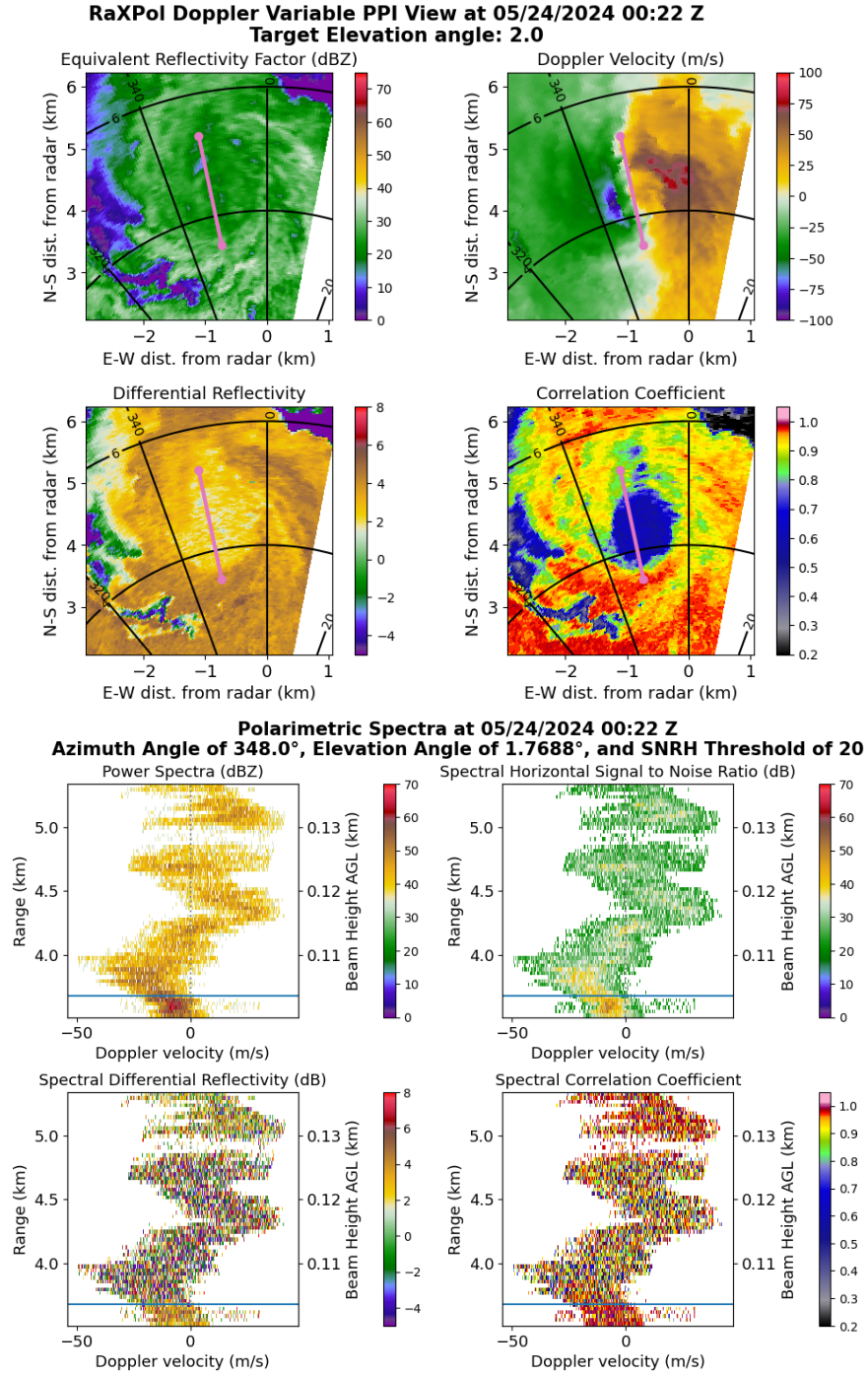


Figure 3.36: RaXPol Doppler variable PPI view at 00:22 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the vortex at a 2° elevation sweep, traversing the main DRVC. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.38.



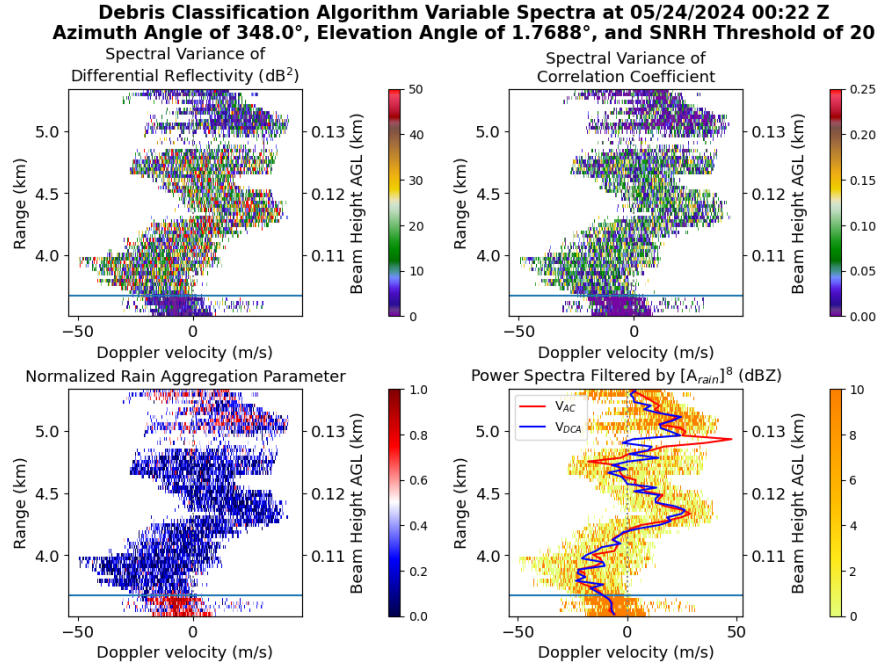


Figure 3.37: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.36 for the Eldorado, OK EF-2 tornado, through the center of the vortex at a 2° elevation sweep, traversing the DRVCs. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.38.

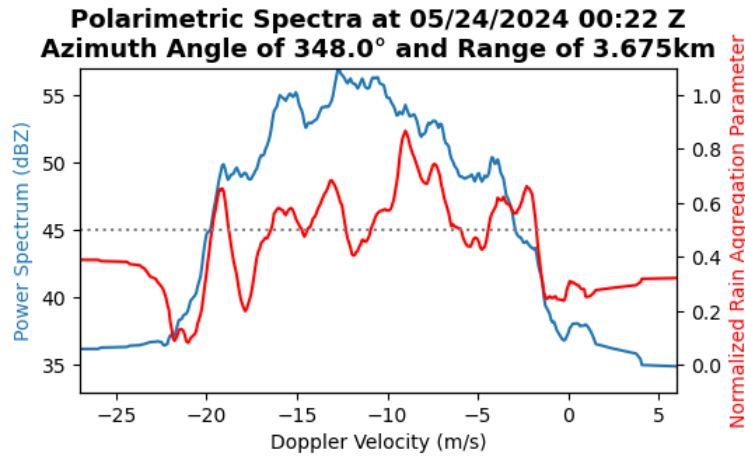


Figure 3.38: Single-dwell spectra for the dwell indicated in Figure 3.36.

In this analysis, a few potential causes can be established. This couplet tends to emerge in areas of strong radial velocity gradients near the RMW of a tornado. It can be co-located with the edge of the debris cloud, but at higher tilts, this co-location becomes less apparent. There are likely multiple factors that contribute to radial velocity gradients and/or differential motion within a resolution volume. The most apparent is debris centrifuging. At low levels ( $0^\circ$  elevation), DRVCs were co-located with DRIs, making differential motion from centrifuging a potential cause. The corner flow region is also a potential mechanism at these tilts, as this area contains significant radial velocity gradients, where radial inflow rapidly changes to vertical and rotational winds.

At the  $1^\circ$  elevation, the DRIs became less co-located with the sharpest change in radial velocity, although they were still contained within the changing radial velocity. At this scan, in Figure 3.35, the bimodal spectrum contained both peaks traveling away from the radar. So, while there were still centrifuging effects, divergence from the tornado center was likely occurring dynamically. We can assume a reasonably high swirl ratio due to the multi-vortex behavior at this time, and as such, downdrafts may have been present, causing real wind divergence. Thus, the DRVC here might be sampling the region above the corner flow region where flow still rapidly changes with radius from the tornado, but outflow is present associated with downdrafts. Away from the surface, these features become less sharp, likely weakening the DRVCs.

### **3.2.3.6 Azimuthal Wind Spectra at Debris-Rain Interface**

So far, only the differential motion in the radial component has been discussed via spectra that traverse the center of a tornado vortex. Numerical simulations also show substantial differences, albeit less significant in the azimuthal wind direction as well.

Bodine (2014) highlights azimuthal differences between sand scatterers and the wind at around  $10 \text{ m s}^{-1}$ .

To find spectra with significant differences between debris and rain, spectra were computed through the DRI at the tornado's wind field edges. This time, we used spectra from sweeps at low levels during the tornado at peak strength. This is because the DRI was slightly removed from the areas of very high spectrum width in the core of the tornado. The spectra that most strongly demonstrate differential motion are shown in Figures 3.39 and 3.40.

The spectra in this figure are similar to those in Figures 3.20 through 3.22. The DCA output in Figure 3.40 does not show significant amounts of debris. The single-dwell spectra at the peak winds (Figure 3.41) show an overall Gaussian PSD with some degree of spectral components at lower velocities classified as debris, with components with higher velocities classified as rain. The average correction of azimuthal speed within the 20 range gates surrounding the peak winds is  $0.46 \text{ m s}^{-1}$ . The highest corrections are on the order of  $5 \text{ m s}^{-1}$ . Thus, the DCA filter depicts these dwells as underestimates on average, which matches with simulations.

Overall, there is a weak signal for the stratification of debris and rain in the azimuthal component. Most spectra through the azimuthal portion of the DRI in this stage do not have significant corrections, and no clear patterns with regard to debris moving at a slower azimuthal velocity. Results for these spectra are inconclusive, though some spectra show corrections in the expected direction.

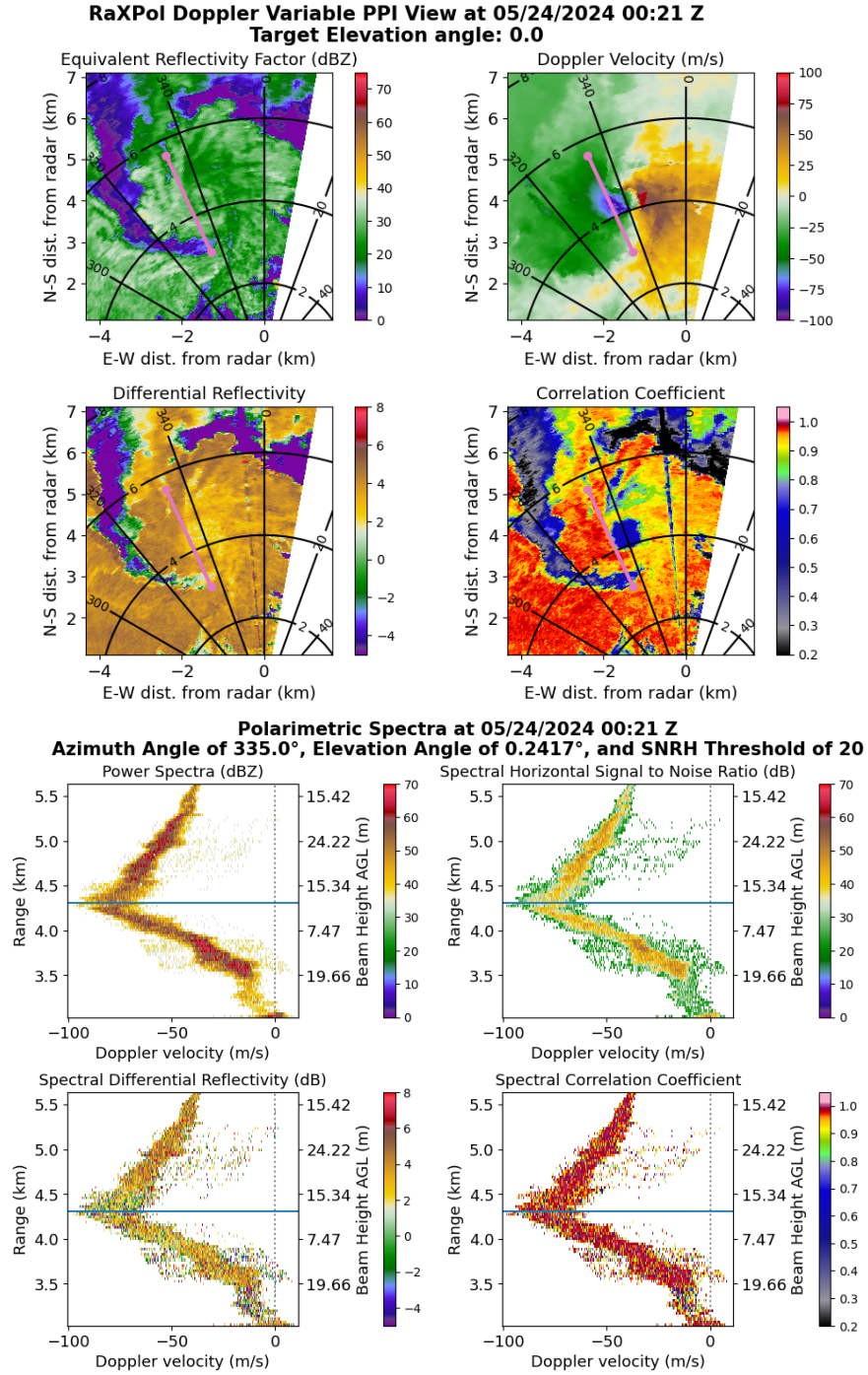


Figure 3.39: RaXPoI Doppler variable PPI view at 00:21 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the edge of the vortex at a  $0^\circ$  elevation sweep, traversing DRI. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.41.

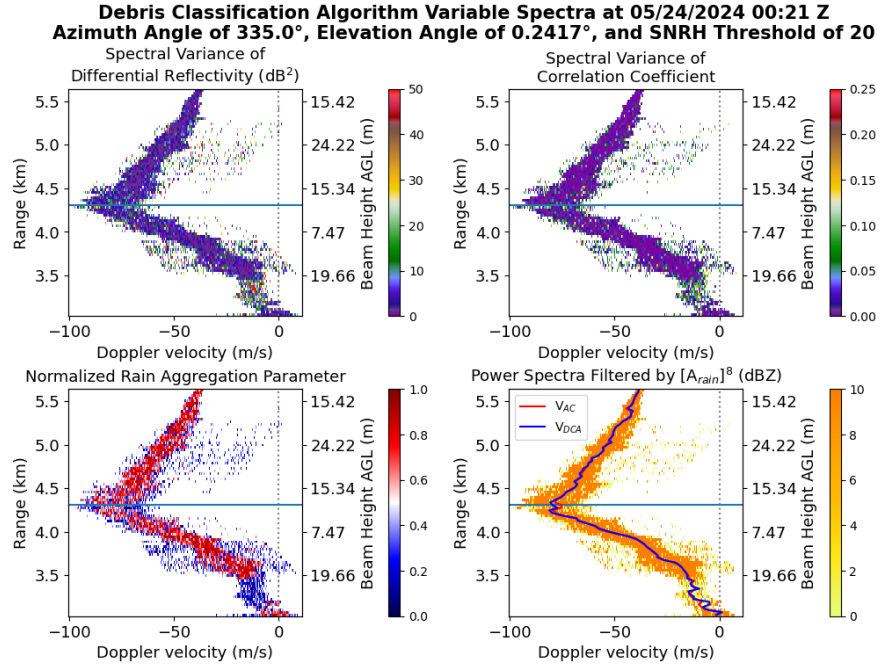


Figure 3.40: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.39 for the Eldorado, OK EF-2 tornado, through the edge of the vortex at a 0° elevation sweep, traversing DRI. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.41.

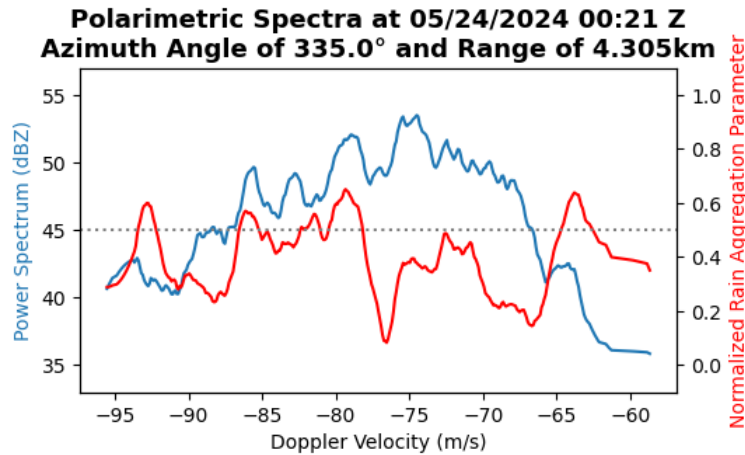


Figure 3.41: Single-dwell spectra for the dwell indicated in Figure 3.39.

## **3.2.4 Stovepipe Phase Spectral Analysis**

### **3.2.4.1 Tornado Progression During Phase**

After crossing the Oklahoma state highway 34 to the north of RaXPol, the tornado became rain-wrapped. It moved out of the multi-vortex phase sector scan area, and the VCP had to be adjusted to scan azimuthal angles to the northeast of the radar. Additionally, as mentioned in Section 2.2.3, a hill resulted in beam blockage and the inability to collect scans of the tornado below  $2^\circ$  elevation.

After the VCP was adjusted, a secondary tornado/satellite vortex was observed by the radar to the north of the original tornado cyclone. This tornado was rated EF-0 and moved south towards the main tornado, presented in Figure 3.42. Eventually, both areas of vorticity merged, and a stovepipe tornado rapidly took shape within seconds, as seen in 3.43. This tornado had a very similar morphology to the satellite vortex observed near the May 31, 2013 El Reno, OK EF-3 (Bluestein et al. 2015).

This phase had a very stable vortex, which became wider as it drifted northeast into forward-flank precipitation and eventually dissipated. The stovepipe phase was defined by a symmetric and laminar vortex structure.

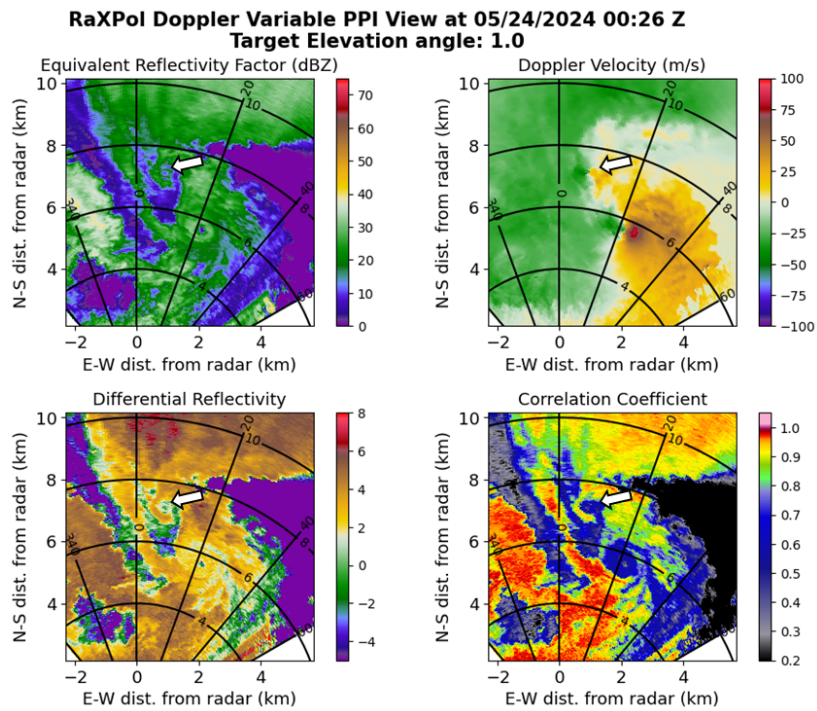


Figure 3.42: RaXPol Doppler variable PPI view at 00:26 UTC. A satellite tornado was observed north of the main tornado cyclone, marked with a white arrow.



Figure 3.43: Photographs by the author of the tornado at 00:29 and 00:31 UTC, showing that the tornado became fully condensed in 2 minutes.



#### 3.2.4.2 Spectra Through Tornado Center

During this phase of the tornado, unfortunately, observations of a DRVC were not possible at any tilt because the beam height was much higher and above the surface layer. The same methodology employed during the multi-vortex phase was followed to compute spectra through the center, with an expectation of similar results to Section 3.2.3.5. A sweep during the peak strength of the tornado in this phase, shortly after the merging of the satellite and main vortices, was examined first.

Due to the distance and elevation angle, the PPI plots in Figure 3.44 contain a three-tier structure. Unlike the multi-vortex phase where the hook was filled with debris, this structure shows a WEH in the middle surrounded by a debris cloud, which is all surrounded by rain. For the spectra through the center, there are clear areas of debris present in the DPSDs and flagged by the DCA in Figure 3.45. However, there are no significant velocity corrections. Because the reflectivities were below the SNRH threshold, the DCA does not provide corrected velocities in the WEH. A single-dwell PSD from within the interface region, in Figure 3.46, exhibits a somewhat Gaussian shape. The corresponding spectrum of  $An_{rain}$  exhibits seemingly random classifications, but towards the PSD peak at  $10 \text{ m s}^{-1}$ , spectral components are classified as rain. For the center of this specific sweep, there is no significant stratification between debris and rain in Doppler space. The tornado velocity field remained in fairly steady state in laminar and symmetric character, and grew wider with time and became increasingly rain wrapped.

The spectra in Figures 3.47 and 3.48 were collected one volume scan later. The tornado had weakened somewhat but had significantly widened into a large stovepipe.

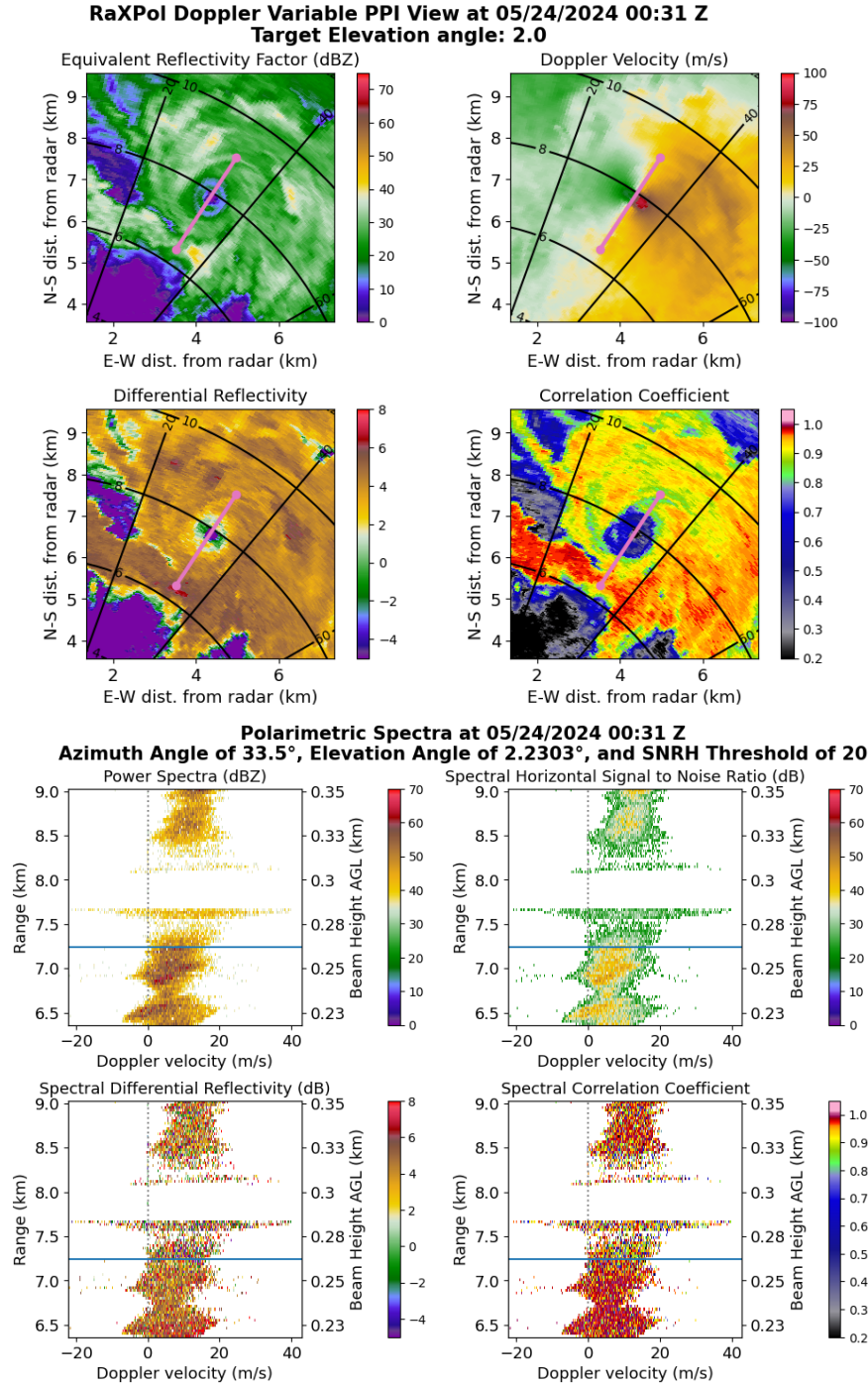


Figure 3.44: RaXPoI Doppler variable PPI view at 00:31 UTC (top 4) and accompanied DPSP range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the vortex at a 2° elevation sweep. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.46.

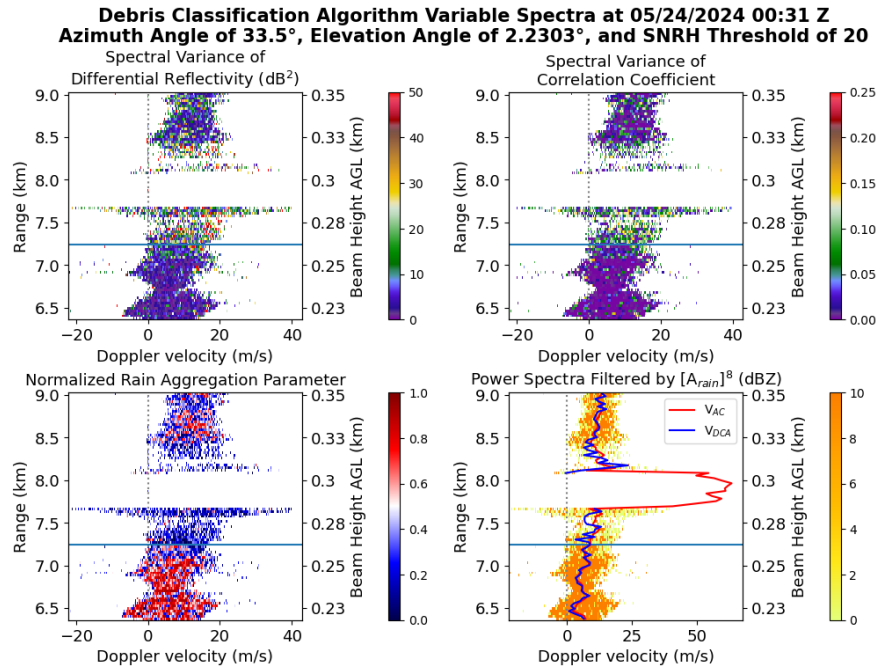


Figure 3.45: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.44 for the Eldorado, OK EF-2 tornado, through the center of the vortex at a 2° elevation sweep. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.46.

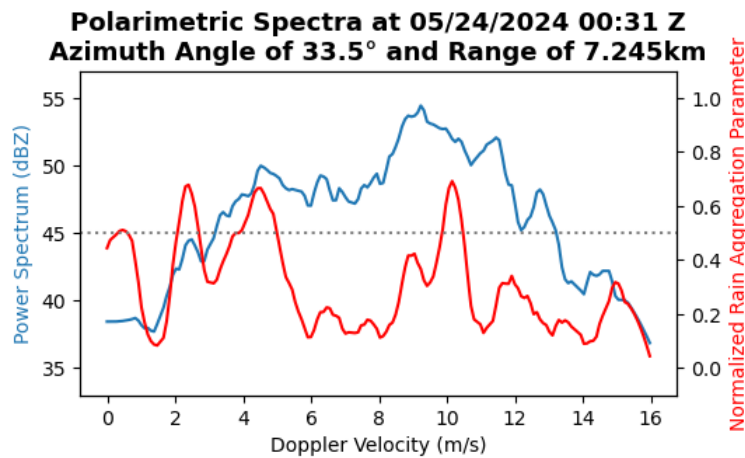


Figure 3.46: Single-dwell spectra for the dwell indicated in Figure 3.44.

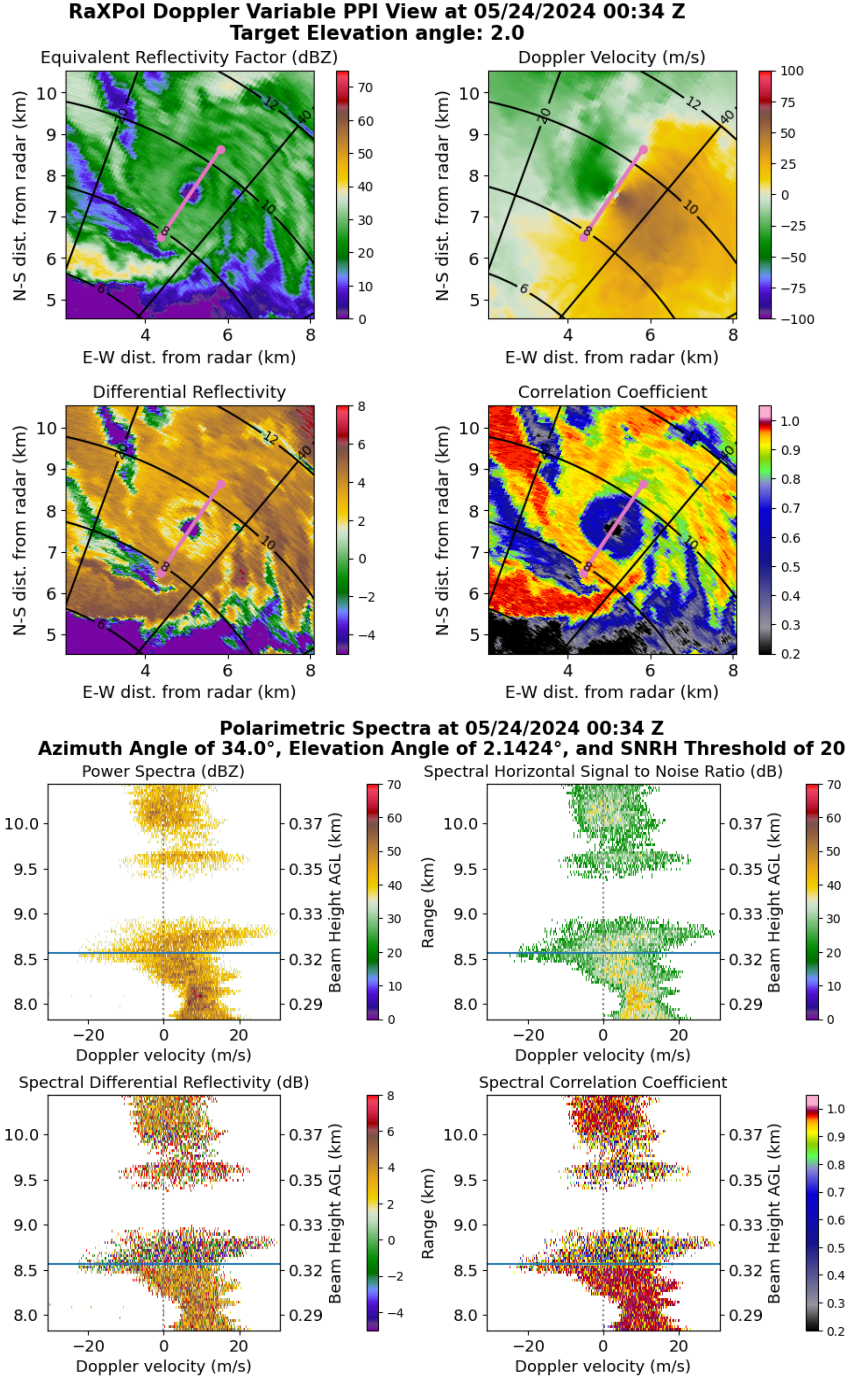


Figure 3.47: RaXPol Doppler variable PPI view at 00:34 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the center of the vortex at a 2° elevation sweep. The blue horizontal line denotes the dwells used for single-dwell spectra computation in Figures 3.49.

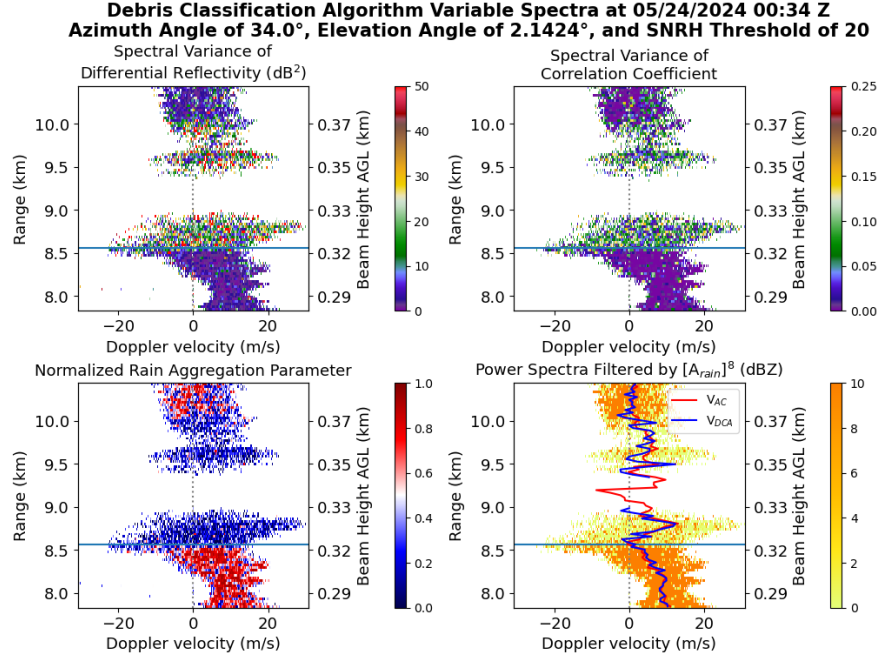


Figure 3.48: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.47 for the Eldorado, OK EF-2 tornado, through the center of the vortex at a  $2^\circ$  elevation sweep. The blue horizontal line denotes the dwells used for single-dwell spectra computation in Figures 3.49.

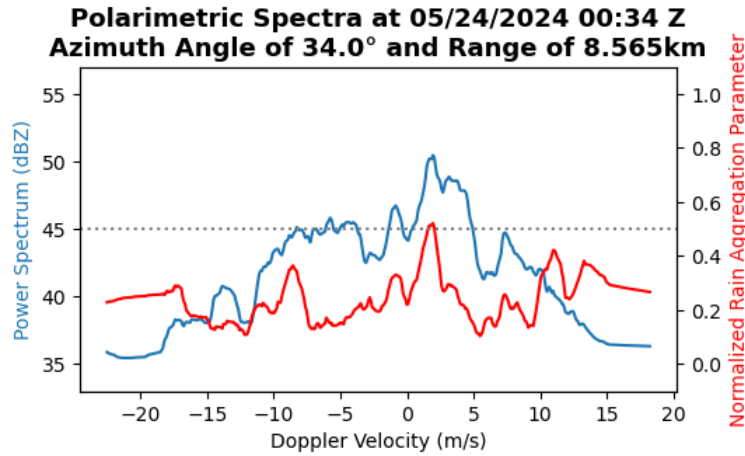


Figure 3.49: Single-dwell spectra for the dwell indicated in Figure 3.47.

The regions of the PSDs adjacent to the weak returns associated with the weak echo hole have a mean positive velocity, owing to the tornado’s motion away from the radar. At this stage, the DRI closer to the radar was tight. The DCA filter-corrected velocity does not result in a large change here, but it does result in a correction in the correct direction, which is towards positive Doppler velocity and decreased radial scatterer divergence. Figure 3.49 shows a bimodal PSD with peak reflectivity barely classified as rain at  $2.5 \text{ m s}^{-1}$ . There is a secondary peak with negative velocity around  $-5 \text{ m s}^{-1}$ . This debris peak represents debris being centrifuged out of the back of the tornado towards the radar as it moves away. The true velocity of the pixel is already positive, thus, the correction is limited, as the filter serves to amplify further the main rain peak in comparison with the secondary debris peak. Even at this elevation, there remained a slight signal of differential motion of debris and rain in the DRI. Additionally, this DRI is narrow. Within the span of 120 m surrounding the gate in Figure 3.49, the spectra become rain-filled with high  $An_{rain}$  or debris-filled with low  $An_{rain}$ , as seen in Figure 3.50.

An additional observation about the PPI plots and spectra in Figure 3.47 is in the differential reflectivity. Particularly, a “half-ring” of elevated differential reflectivity appears to be present within the area of depressed correlation coefficient, meaning that this signature is likely debris even though it has differential reflectivity values as high as rain. The cause of this is unknown, but could potentially be attributed to debris orientations becoming uniform in processes similar to those described in Umeyama et al. (2018).

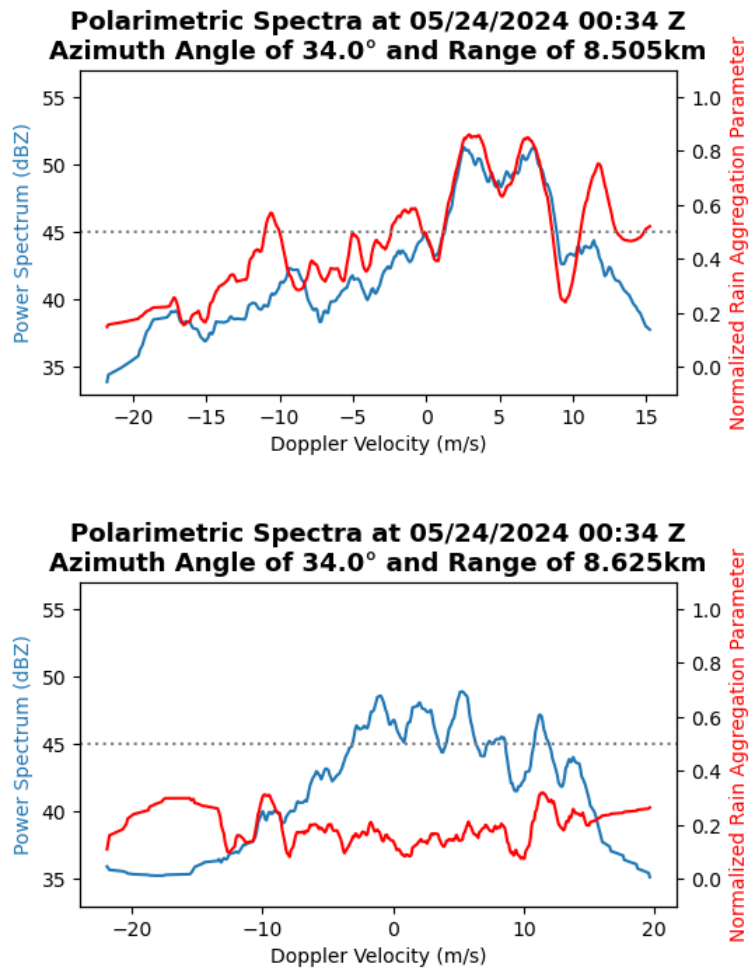


Figure 3.50: Single-dwell spectra in the two range-adjacent dwells in Figure 3.49. The top spectrum (2 range gates closer to the radar) contains a peak with mainly rain, while the bottom spectrum (1 range gate further from the radar) contains a peak with mainly debris. These two figures are 120 m from each other.

### 3.2.4.3 Spectra Through Tornado Edge

With the DRI narrow at this time due to rain wrapping, the same methodology from Section 3.2.3.6 was applied to collect spectra at the edge of the DRI to find differential azimuthal motions. The analysis begins with spectra through the stovepipe phase at peak strength.

Figures 3.51 and 3.52 depict the laminar nature of the tornado. Spectrum widths within this region are narrow, suggesting limited turbulence. The DCA has some spectral components at the velocity peak at 8 km flagged as debris, but the corrections are not significant.

Figure 3.53 shows a fairly narrow spectrum, with the X-axis ranging from 50 m s<sup>-1</sup> to 65 m s<sup>-1</sup>. On average, the slower velocity spectral components are classified as rain, but the PSD shows that these velocities have a higher reflectivity peak. Thus, the velocity correction is minimal. In this sweep, spectral components flagged as debris are lower reflectivity and non-peak areas, and thus the DCA does not significantly correct them. The average correction within the 20 range gates surrounding the peak winds is -0.18 m s<sup>-1</sup>. Differential motions are not significantly observed.

Later in the tornado's life, the tornado exhibited stronger differential motions in the azimuthal component. Spectra computed in the DRI revealed substantial corrections in the azimuthal component.

In Figures 3.54 and 3.55, spectra from the northwest DRI have significant stratification between debris and rain. It is seen in the  $An_{rain}$  spectra that rain has a faster azimuthal speed than debris through the peak intensity region of the tornado, and the PSDs show that this region is not associated with a maximum in reflectivity. At



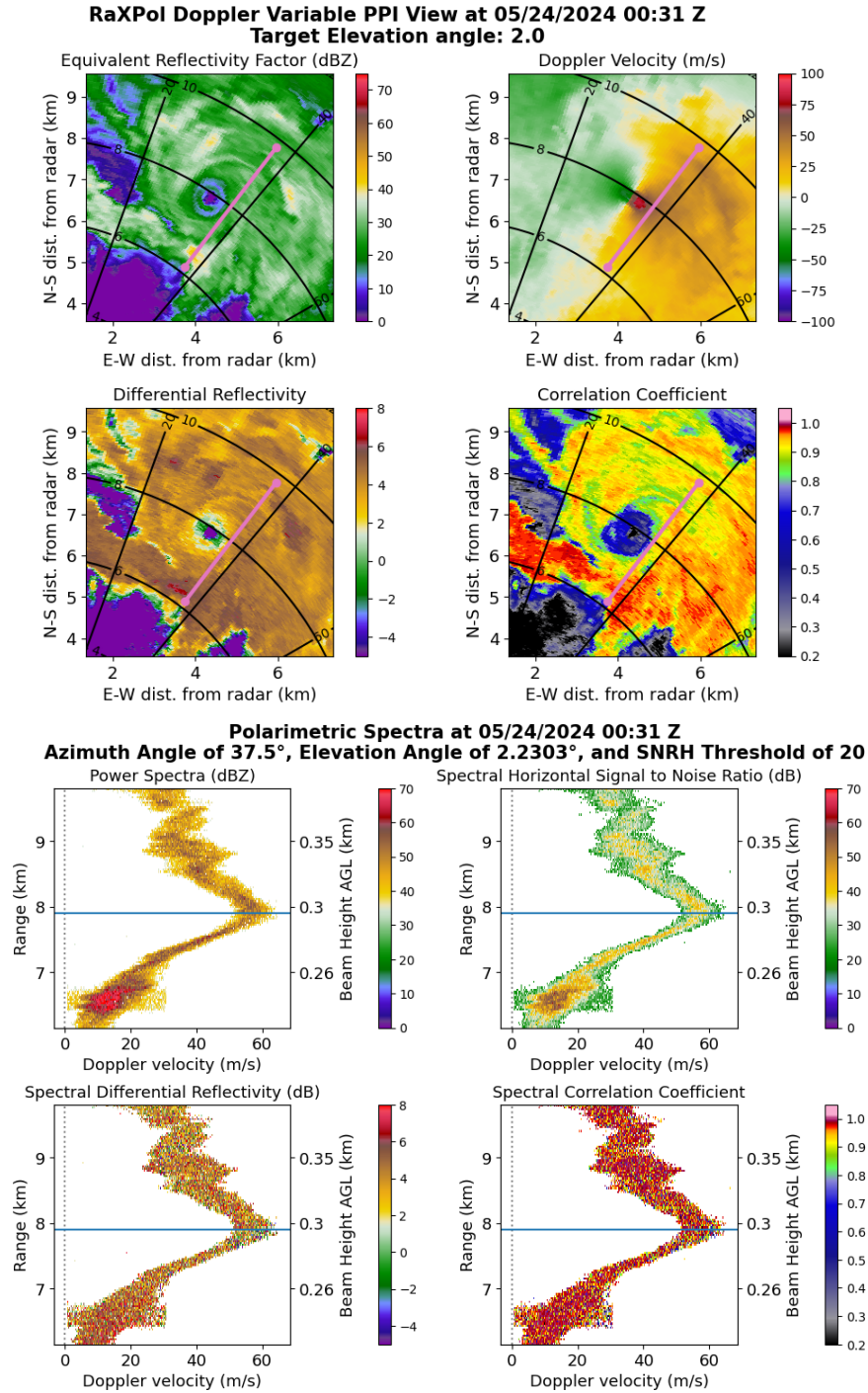


Figure 3.51: RaXPol Doppler variable PPI view at 00:31 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the southeast edge of the vortex, through the DRI, at a 2° elevation sweep. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.53.

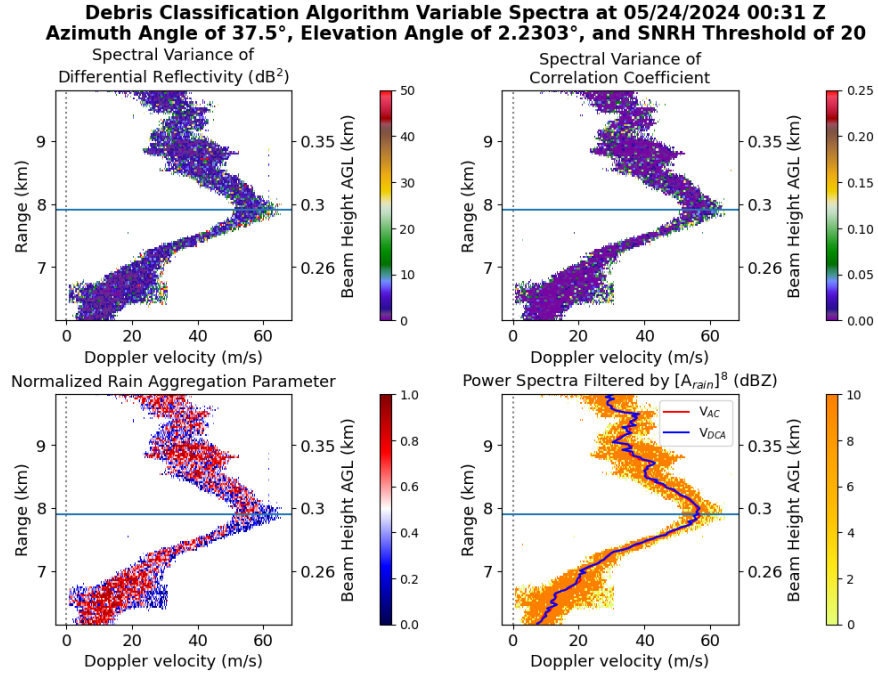


Figure 3.52: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.51 for the Eldorado, OK EF-2 tornado, through the edge of the vortex, through the DRI, at a 2° elevation sweep. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.53.

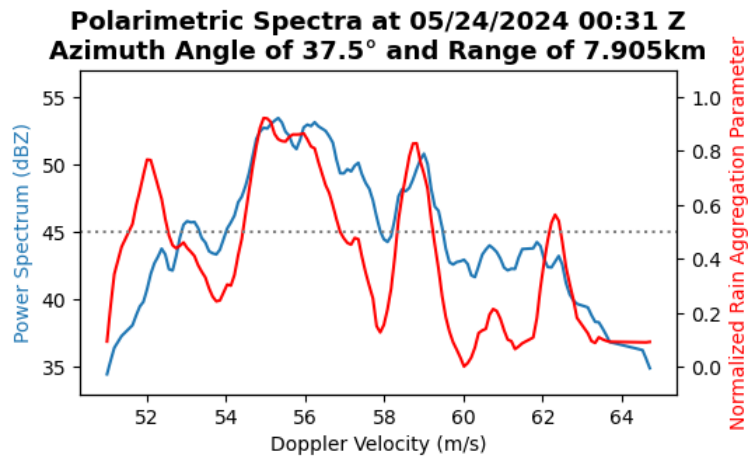


Figure 3.53: Single-dwell spectra for the dwell indicated in Figure 3.51.

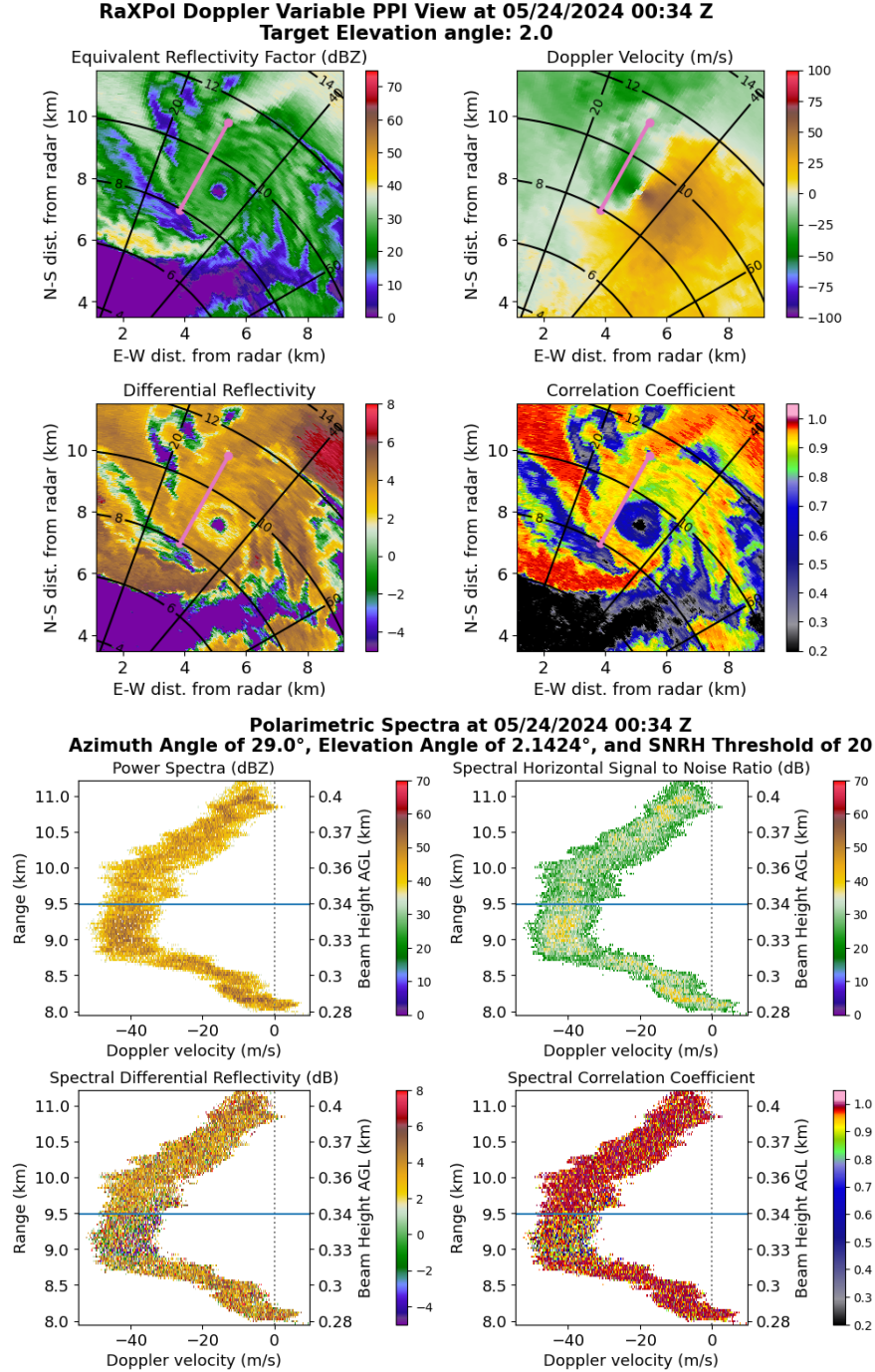


Figure 3.54: RaXPol Doppler variable PPI view at 00:34 UTC (top 4) and accompanied DPSD range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the northwest edge of the vortex, through the DRI, at a 2° elevation sweep. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.56.

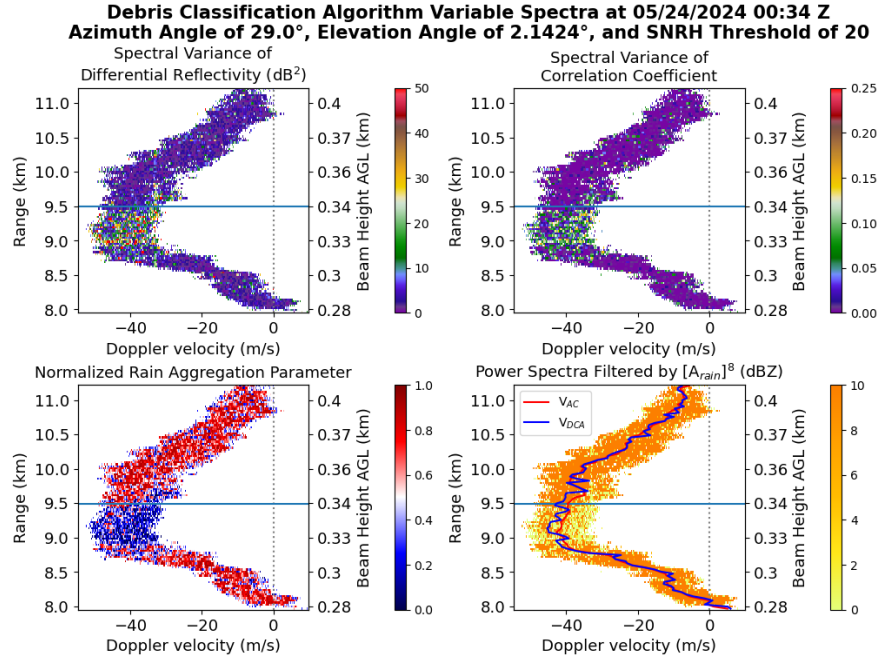


Figure 3.55: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.54 for the Eldorado, OK EF-2 tornado, through the northwest edge of the vortex, through the DRI, at a 2° elevation sweep later in the life of the stovepipe phase of the tornado. The blue horizontal line denotes the dwell used for single-dwell spectra computation in Figure 3.56.

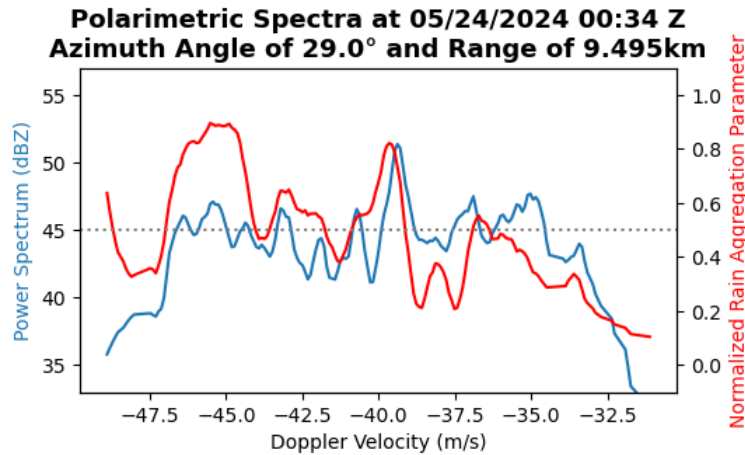


Figure 3.56: Single-dwell spectra for the dwell indicated in Figure 3.54.

this location, debris was clearly traveling more slowly than rain, and the DCA filter corrects these velocities. The average correction within the 20 range gates surrounding the peak winds is  $-1.83 \text{ m s}^{-1}$ . The highest corrections are on the order of  $-5 \text{ m s}^{-1}$ .

A single-dwell PSD taken through the region of the correction in Figure 3.56 shows an un-skewed, wide main peak. The faster azimuthal speeds are clearly classified as debris. The filter successfully corrects velocities by amplifying faster spectral components.

On the southeast edge of the tornado in Figures 3.57 and 3.58, spectra computed are similar to those on the northwest edge. Here, there is similar stratification, where it is seen in the  $An_{rain}$  spectra that spectral components with higher speed are more likely to be rain. Debris spectral components also do not appear to have significantly lowered PSD components. Unlike on the northwest side, however, the DCA filter does not correct these velocities significantly. Doppler velocities are slightly corrected towards increased tangential velocity, and the average correction within the 20 range gates surrounding the peak winds is  $0.18 \text{ m s}^{-1}$ . Although the spectra are qualitatively stratified, this does not result in a significant wind-field correction.

Overall, evidence of stratification between debris and rain appeared in some dwells, but remained inconsistent. Significant corrections of velocity by the DCA filter typically increased azimuthal velocity.



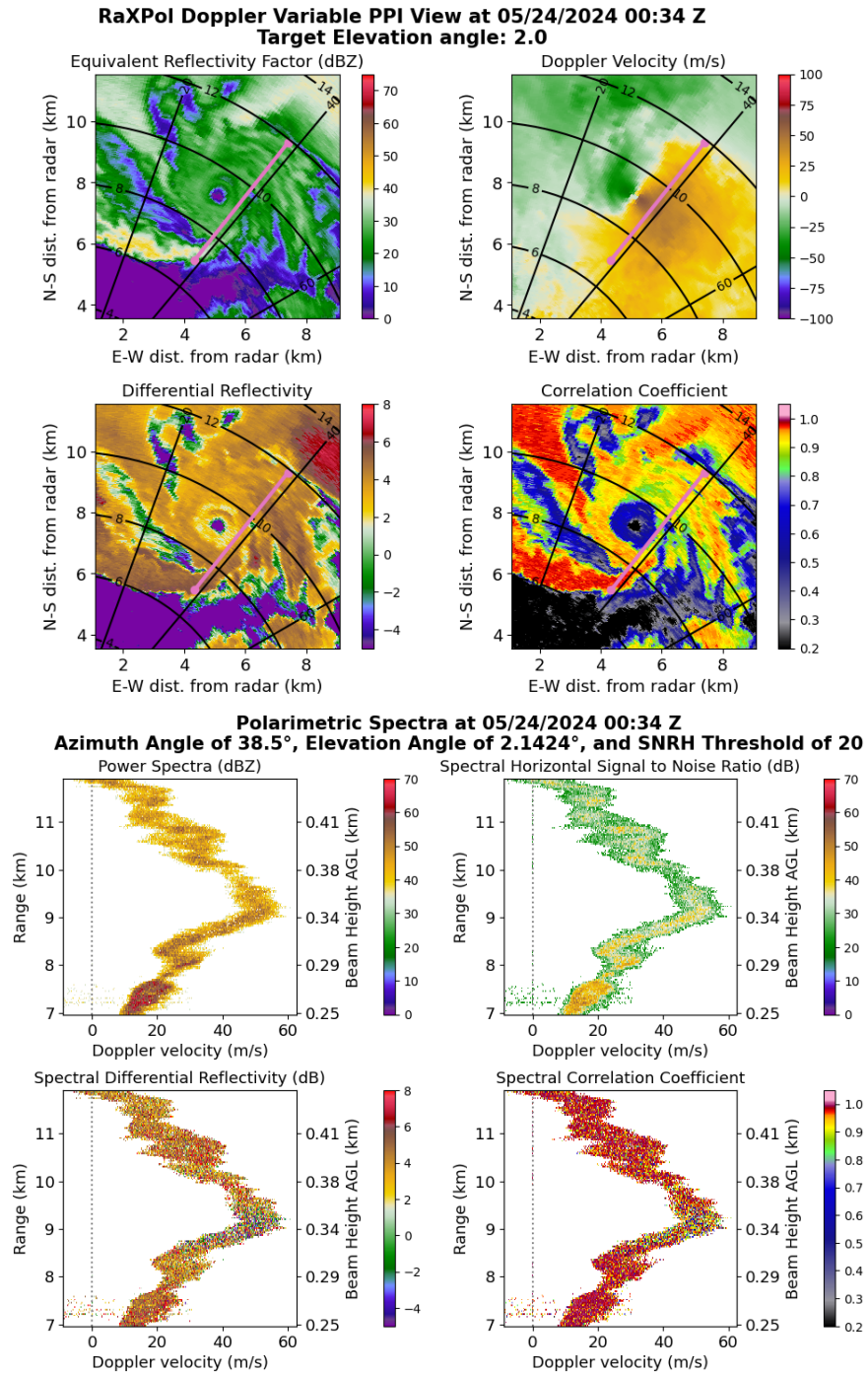


Figure 3.57: RaXPol Doppler variable PPI view at 00:34 UTC (top 4) and accompanied DPSP range-Doppler plot (bottom 4, see Section 2.3.2.1) for the Eldorado, OK EF-2 tornado, through the southeast edge of the vortex, through the DRI, at a 2° elevation sweep.

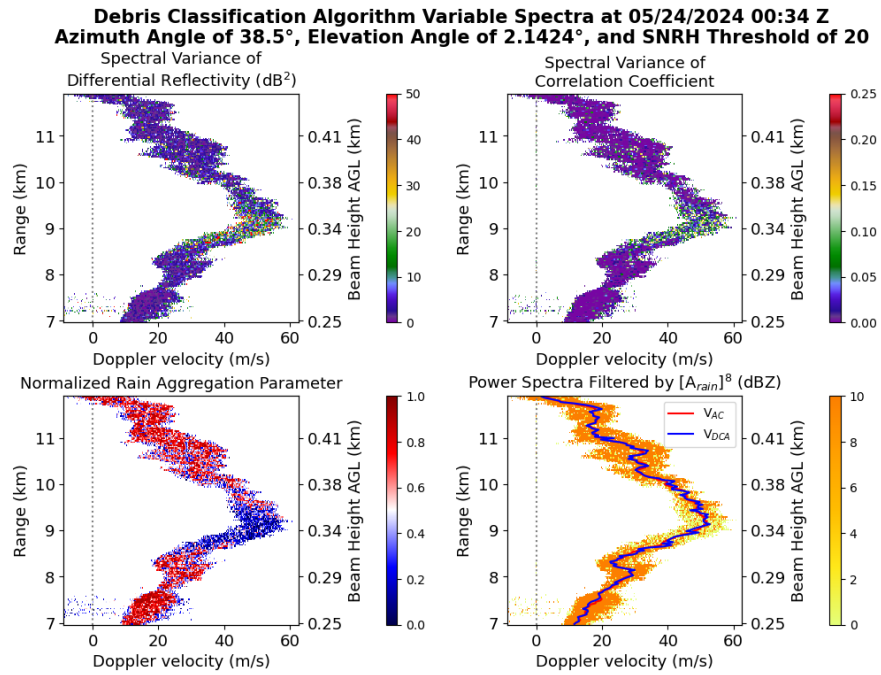


Figure 3.58: DCA variable spectra in range-Doppler plots (see Section 2.3.2.2) computed from spectral estimates in Figure 3.57 for the Eldorado, OK EF-2 tornado, through the southeast edge of the vortex, through the DRI, at a 2° elevation sweep.

## Chapter 4

### Conclusions

#### 4.1 Summary

Dual-polarization spectral densities serve as an important tool in correcting velocity estimates. They can distill components of a Doppler spectrum by their polarimetric characteristics, like uniformity ( $s\tilde{\rho}_{HV}$ ) or aspect ratio ( $s\tilde{Z}_{DR}$ ). In tornadoes, debris has a very high spectral variance, which can be exploited to classify spectral components as debris or rain. DCA outputs were useful in aggregating contributions of different DPSD variables, like  $s\tilde{Z}_{DR}$  and  $s\tilde{\rho}_{HV}$ , into a single output with straightforward interpretation,  $An_{rain}$ . Values of  $An_{rain}$  above 0.5 were likely to be classified as rain, and below 0.5 were likely to be classified as debris.

The convergence of DPSDs requires a scanning strategy specific to maximizing frequency resolution to facilitate the convergence of spectral components. Unfortunately for dish-based radars, this means sacrificing temporal resolution. However, if the separation between debris and rain is significant, even spectra that have not converged are capable of showing this difference at a relatively low frequency resolution. Additionally, as a corollary to convergence with increasing frequency resolution, high spectral variance of DPSDs is potentially a physical characteristic of debris returns in tornadoes, which is exploited by the DCA.

The tornado displayed many stages and structures, allowing the collection of a varied dataset. Before spectral analysis was conducted, when the radar was scanning at



low levels during the multi-vortex phase, it was already possible to ascertain the effects of centrifuging in the DRVC regions. These regions manifested as sharp changes in velocity with range and anomalously high winds radially outwards from the tornado, often occurring squarely at the DRI. These winds were likely driven by debris centrifuging and the naturally sharp wind gradients in the DRVC regions of a tornado, both in and above the corner flow region. Surrounding these regions was generally strong radial inflow, able to be sampled due to the close range to the radar and low elevation angle of the scan detecting ground clutter nearby. These observations are supported by numerical simulations (Lewellen 1993; Dowell et al. 2005; Fischer and Dahl 2022; Rotunno and Bluestein 2024) and photographic evidence (Dowell et al. 2005).

The polarimetric spectral analysis of the tornado provided evidence of differential motion of rain and debris at the DRI. This was especially the case with resolution volumes sampling radial winds within the DRVC regions of the tornado. Power spectra in these dwells ranged from significantly bimodal to wide monomodal. The radially inbound Doppler velocity components had DPSD characteristics of rain, and radially outbound Doppler velocity components had DPSD characteristics of debris. The DCA aggregation filter was able to correct aggressively for the differential motion of debris in the DRVC regions, correcting towards the motion of rain, which is likely to be closer to the true velocity of the wind due to its smaller mass and size. The differential motion was likely caused by two factors: centrifuging and vertical wind shear between the shallow inflow layer near the ground and debris centrifuging slightly aloft.

In other areas of the tornado, like aloft above the DRVC regions and at the edges of the tornado sampling purely azimuthal winds, the signal for differential motion was much weaker. At these locations, scatterers depend only on differential centrifuging to achieve differential motion rather than being aided by wind shear in the resolution volume. Above the DRVC regions, the shallow inflow layer is no longer sampled. Thus,

in the radial component, winds that may be purely azimuthal are in the same resolution volume as smaller or lighter debris, which was lofted higher in altitude. Differential motions become lower here as the degree of centrifuging decreases and the true radial wind component lessens, as well as heavier debris being less common and falling out at lower altitudes. Weak differential motion was observed in some scans at the DRI in both the multi-vortex and stovepipe phases through the center of the tornado. Power spectra in these phases and locations were more likely to be monomodal, and if there was stratification between debris and rain, it was represented by skewness in  $An_{rain}$ . Similarly, at the edges of the tornado sampling azimuthal winds, the stratification was also inconsistent and minor. When there was a significant velocity correction, however, it tended to be in the correct direction. This was likely due to similar reasons of small debris sizes and reliance on only centrifuging, without the added boost of wind shear within the resolution volume, and remained true for both the multi-vortex phase and the stovepipe phase.

## 4.2 Future Work

Significant opportunities for future work are possible, both in the field and out. The first possibility involves conducting simulations with tornado and debris properties similar to the ones observed in the Eldorado case. The closest existing simulations are the sand simulations from Bodine (2014). Simulations with dust and grass combined for wind speeds similar to those exhibited in the tornado would enable direct comparison between the observed data and simulations, which was not possible with this dataset. This type of simulation could also spur the development of different membership functions for a new DCA. The DCA in Schneider et al. (2025), although very successful in identifying debris spectral components in the observed tornado, was originally designed

for wood boards. It would be useful to follow a similar methodology and design a DCA accurate to this tornado and potentially tornadoes with many more debris types based on idealized simulations. Additionally, the DCA and other spectral analysis could also be augmented by examining spectral differential phase, another polarimetric radar variable which compares the phase difference between horizontally and vertically polarized returns. Other, simpler variables could also be used in a DCA, like reflectivity, which can be higher for larger or more reflective debris.

Unfortunately, the DCA aggregation filter's radar-centric approach is limited because most dwells lack comparable amounts of debris and rain. This is especially true in the core of the tornado, where no substantial rain signal exists for correction, and thus, correction magnitudes tend to be lower and introduce noise into the velocity field. Nolan (2013) presented a methodology to add the effects of centrifuging to a derived wind field, where a wind bias is selected, scaled to the observed centripetal acceleration at a given radius, and then applied to a wind field. This centrifuging bias resulted in strong downdrafts within the vortex center. Similarly, if a wind field is derived from radar data via multiple-Doppler analysis or GBVTD, these centrifuging biases are already incorporated. An alternative, more physical method to correct the velocity field can be implemented as the opposite of the above technique. Spectra, particularly those significantly corrected by the DCA aggregation filter, can be used to give estimates of peak radial centrifuging, and these can be scaled based on the centripetal acceleration to estimate corrections in locations where both types of scatterers are not present. This type of correction is another future avenue that can be explored.

In the field, fortunately, this tornado stayed mainly over rural land, but it is also important to determine what occurs when a tornado lofts larger debris, which may be structural, like wood planks or bricks. These objects are likely to centrifuge at a much higher degree, thanks to their increased mass, and may result in larger velocity biases.

It is important to study the degree of velocity bias as it compares to that hypothesized in simulations.

Finally, the advent of digitally scanning radars, like Horus (Palmer et al. 2023), PAIR (polarimetric atmospheric imaging radar, Salazar et al. 2019), or other dual-polarization phased array radars, enable spectral analysis that may not be feasible with traditional dish radar. As discussed in Section 3.1, for reasonable convergence of spectral components, the scanning strategy employed is required to result in sufficient Doppler resolution. With a dish radar, slow scanning across an entire sweep was required for spectral computations in the small area of a tornado. With digital scanning, the beam can be focused in on the area of interest, transmitting enough pulses in the localized area near the tornado to collect spectral estimates. Since tornadoes are spatially limited, temporal resolution would not significantly be lost in this process.

Overall, many possibilities exist for improving velocity estimates. Evidence of centrifuging effects in real tornadoes appears in the computed spectra. Accurate velocity estimates are crucial for studying tornado dynamics, as small errors can significantly contaminate important kinematic retrievals like vertical wind. Implementing spectrally aware velocity corrections may enable more accurate observational studies of tornado dynamics.

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